The Seismotectonics of the Po Plain (Northern Italy): Tectonic Diversity in a Blind Faulting Domain

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Abstract—We present a systematic and updated overview of a seismotectonic model for the Po Plain (northern Italy). This flat and apparently quiet tectonic domain is, in fact, rather active as it comprises the shortened foreland and foredeep of both the Southern Alps and the Northern Apennines. Assessing its seismic hazard is crucial due to the concentration of population, industrial activities, and critical infrastructures, but it is also complicated because (a) the region is geologically very diverse, and (b) nearly all potential seismogenic faults are buried beneath a thick blanket of Pliocene–Pleistocene sediments, and thus can be investigated only indirectly. Identifying and parameterizing the potential seismogenic faults of the Po Plain requires proper consideration of their depth, geometry, kinematics, earthquake potential and location with respect to the two confronting orogens. To this end, we subdivided them into four main, homogeneous groups. Over the past 15 years we developed new strategies for coping with this diversity, resorting to different data and modeling approaches as required by each individual fault group. The most significant faults occur beneath the thrust fronts of the Ferrara-Romagna and Emilia arcs, which correspond to the most advanced and buried portions of the Northern Apennines and were the locus of the destructive May 2012 earthquake sequence. The largest known Po Plain earthquake, however, occurred on an elusive reactivated fault cutting the Alpine foreland south of Verona. Significant earthquakes are expected to be generated also by a set of transverse structures segmenting the thrust system, and by the deeper ramps of the Apennines thrusts. The new dataset is intended to be included in the next version of the Database of Individual Seismogenic Sources (DISS; http://diss.rm.ingv.it/diss/, version 3.2.0, developed and maintained by INGV) to improve completeness of potential sources for seismic hazard assessment.

Key words: Blind faulting, Seismogenic source, Active tectonics, Seismic hazard, Inherited faults, Po Plain.

1. Introduction

The Po Plain is one of Italy's major geographical features, extending for over 400 km in a roughly east–west direction from the Alps to the Adriatic Sea and corresponding to the drainage basin of the Po River, the country's longest and most important watercourse. Geologically, the Po Plain and part of the current Adriatic Sea comprise the foreland of two opposing, verging fold-and-thrust belts: the Southern Alps to the north, and the Northern Apennines to the south. Although it is underlain by actively growing orogens in the context of ongoing convergence, the Po Plain is an almost perfectly flat area that may appear tectonically quiet, bounded by two mountain ranges. The morphological borders of the Po Plain correspond with the contact between the Quaternary alluvium outcropping in the plain and the pre-Quaternary rocks exposed along the two mountain fronts.

Despite the appearances, however, the Po Plain is in itself quite active. Several earthquakes have hit the plain and both mountain fronts through history (ISIDe Working Group 2010; Rovida *et al.* 2011), testifying that the geodynamic processes which led to the convergence between the two orogens are still active. Because of the occurrence of many historical centers combined with an extraordinary concentration of critical facilities and the associated lifelines, the Po Plain certainly represents a major concern for local administrations and civil defence authorities. Despite these circumstances, and probably due to the scattered nature of local seismicity, much of the Po Plain has entered the Italian seismic code only recently, in 2004, and the new provisions have become mandatory even more recently, in 2010. For these reasons, even earthquakes in the magnitude range 5–6 can be catastrophic, as demonstrated by the 2012 Emilia sequence (M_L 5.9; ISIDe Working Group 2010), which caused 27 casualties and severe damage to monumental and industrial buildings.

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The Italian earthquake record is normally considered complete for M 6+ earthquakes only for the past ca. 700 years (STUCCHI et al. 2004), but major known earthquakes probably recur every 1,000–2,000 years (VALENSISE and PANTOSTI 2001a; GALLI et al. 2008). Consequently, faults that released their most recent large earthquake before 700 years ago may not appear in the historical record. Identifying all potential seismogenic faults and estimating their potential is fundamental for characterizing input for probabilistic seismic hazard assessment, but given the short length of the historical record and the poor surface expression of potential seismogenic sources, we may be surprised by future damaging earthquakes (MULARGIA 2013), a fact that was illustrated by the 2012 Emilia sequence. The 20 and 29 May earthquakes occurred in an area for which no historical seismicity was reported, but where seismogenic sources had been identified through tectonic geomorphology and seismic profiles [Database of Individual Seismogenic Sources (DISS) 2.0, VALENSISE and PANTOSTI 2001b; CIUCCI et al. 2002]. Resorting to the geological record is thus mandatory for outlining all seismogenic sources of a problematic area such as the Po Plain and assessing its potential effectively (e.g. FACCIOLI 2013).

Fault plane solutions of major instrumental earthquakes show that compressional mechanisms are dominant in the Po Plain (PONDRELLI et al. 2006; SCOGNAMIGLIO et al. 2012, among others), which agrees with geologic and modern strain (GPS) evidence (e.g. DEVOTI et al. 2011) and with predictions from geodynamic models. Relatively recent active tectonic studies have acknowledged that faulting is dominantly blind and that strain rates are rather low (see VALENSISE and PANTOSTI 2001b; BURRATO et al. 2003; BASILI et al. 2008; TOSCANI et al. 2009; DISS WORKING GROUP 2010; MAESANO et al. 2013; submitted). However, correlating individual seismogenic sources (ISS) and associating them with specific historical earthquakes is not straightforward. A number of additional circumstances add to the complications, including (1) the scattered distribution of seismicity; (2) the large sedimentary rates in the Po Plain proper, often larger than tectonic rates; (3) the presence of inherited landscape and lithological discontinuities in the foothills; (4) locally large human-induced vertical ground motions; and (5) widespread anthropogenic modifications of the natural fluvial network (e.g. the construction of levees and of artificial channels).

Because of these circumstances, for many years, few investigations of active faults in the subsurface of the Po Plain were begun as most field geologists concentrated on the more visible—and thus more scientifically appealing—faults exposed across the Alps and the Apennines. Identifying and characterizing seismogenic sources in the Po Plain requires an integrated approach that combines morphotectonic analyses, seismicity data and subsurface geological data such as commercial seismic lines, wells, in situ stress measurements. Anomalous stream morphology has proven to be the landscape feature that is most sensitive to subtle topographic changes caused by underlying active faults in the flattest portions of the Po Plain, and has been a keystone of our approach (BURRATO et al. 2003, 2012). In the foothill areas where regional uplift dominates, active thrusting is normally identified by differentially uplifted and tilted Quaternary markers, such as river and even coastal terraces (e.g. AMOROSI et al. 1996, 2004; VANNOLI et al. 2004; PICOTTI and PAZZAGLIA 2008; PONZA et al. 2010; GUNDERSON et al. 2013, 2014).

The case of the Po Plain is especially complex, however, also in view of its extreme tectonic diversity, evidenced by the wide range of hypocentral depths and focal mechanisms displayed by current seismicity. The characteristic flat-and-ramp structure and arcuate shape of most compressional systems worldwide, including those encircling the Po Plain, imply that active thrust and reverse faults occur at different depths and with different strike, thus requiring multiple and flexible geological approaches and a wealth of subsurface data. The presence of inherited faults in the basement rocks of the Po Plain further complicates the matter, as their reactivation under the current compressional regime may result in earthquakes that are difficult to anticipate based on the evolution of the overlying mountain chains.

We present an updated overview of the seismogenic sources of the Po Plain and of the
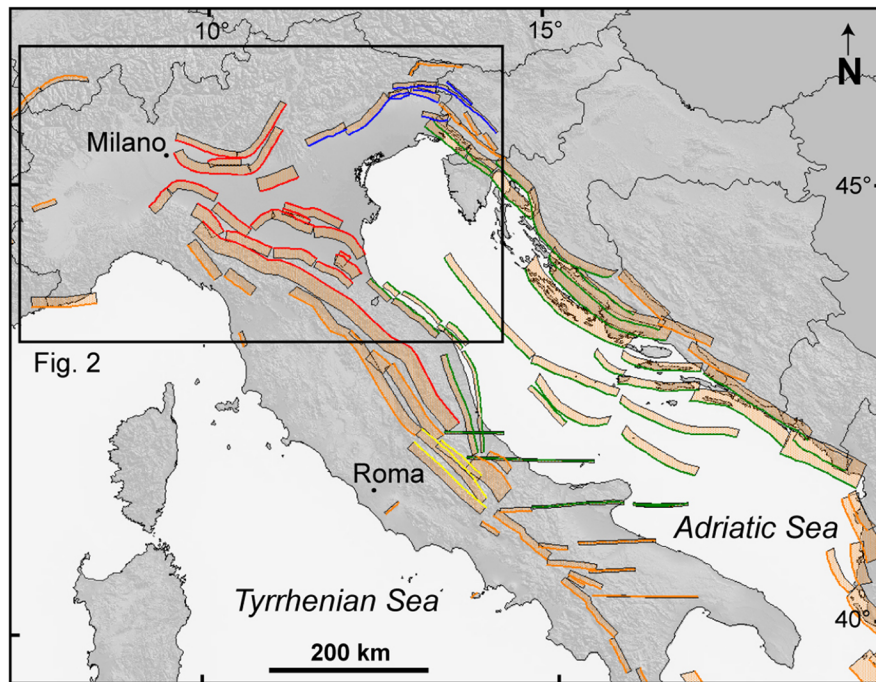
Figure 1

Composite seismogenic sources of the DISS database. Sources of the Po Plain are shown in *red* and described in this work; sources of northeastern Italy and western Slovenia are shown in *blue* and described in Burrato *et al.* (2008); sources of the Adriatic Domain are shown in *green* and described in Kastelic *et al.* (2013); extensional sources of the Central Apennines are shown in *yellow* and described in Vannoli *et al.* (2012); other sources in *orange*

two adjacent mountain fronts, most of which are listed in the latest version of the DISS database (Fig. 1), focusing specifically on their existence, geometry, kinematics, earthquake potential and complexity. Their characterization in terms of slip rate and seismogenic potential is required for probabilistic seismic hazard analysis but it is beyond our current scopes. For a summary of strain rates across the Po Plain and beyond the reader may refer to Petricca *et al.* (2013) and Maesano *et al.* (submitted).

All potential seismogenic sources correspond to blind faults, which are based on scientific inferences and subject to considerable uncertainties, partially compensated by the exceptional density of underground information. We will discuss the reasoning adopted in each single case, the uncertainties, and the research directions that may help setting firmer constrains during future investigation efforts (e.g. Faccioli 2013).

2. Geological and Seismological Framework: Data and Interpretations

2.1. Regional Tectonic Framework

The Po Plain is the foreland of two fold-and-thrust belts verging toward one another: the S-verging central Southern Alps to the north, and the N–NE-verging Northern Apennines to the south. The two belts developed due to convergence between the African and European plates beginning in the Cretaceous, and were associated with two oppositely dipping subduction zones involving both European and African lithosphere (Carminati and Doglioni 2012, and references therein). Thus, the Po Plain coincides with the foreland of both chains; it also hosts their foredeeps, and for this reason it has undergone steady subsidence and sedimentation over the past 5 Ma (Fantoni and Franciosi 2010).

Buried compressional structures comprising the most advanced fronts of the two orogens were extensively explored by seismic reflection lines and deep well logs obtained for oil exploration after the end of World War II. The local Pliocene–Quaternary sedimentary sequence is characterized by uneven thickness, ranging between several thousand meters in the depocenters to a hundred meters over the top of the buried anticlines (BIGI *et al*. 1992; GHIELMI *et al*. 2013). Tectonically induced, short-wavelength variations in the geometry of the sedimentary sequence are superimposed onto a longer-wavelength signal due to the larger subsidence of the Northern Apennines foredeep with respect to that of the Southern Alps. As a result, the thickness of the sedimentary wedge—and similarly the depth and dip of the basal detachment of the thrust wedge—generally increases towards the Northern Apennines mountain front (i.e., southward; MARIOTTI and DOGLIONI 2000; BIGI *et al*. 2003). Rapid asymmetric subsidence caused by the tectonic load of the Northern Apennines chain across the whole Po Plain created a large accommodation space, resulting in large sedimentation rates in the complex system of elongate foredeeps (e.g. BARTOLINI *et al*. 1996; GHIELMI *et al*. 2013). Subsidence takes place at up to 2–3 mm/a (ARCA and BERETTA 1985; CARMINATI and MARTINELLI 2002; CUFFARO *et al*. 2010) and has always been consistently larger than the tectonic uplift rates along the active thrust-related folds, which caused them to drown in the subsurface of the Po Plain.

The outer arcs of the Northern Apennines are currently buried below the Pliocene–Quaternary marine and continental deposits infilling the Po Plain basin and are organized in three complex fold systems: the Monferrato Arc, Emilia Arc, and Ferrara-Romagna Arc, respectively from west to east (Fig. 2). To the rear of the Emilia and Ferrara-Romagna arcs, the subemergent Pedeapenninic Thrust Front (BOCCALETTI *et al*. 1985) lies at the foot of the Northern Apennines. This important morphological and structural feature has been interpreted in recent studies also as the anticline forelimb of a deeper thrust ramp, and thus not as evidence of a shallow thrust (PICOTTI and PAZZAGLIA 2008). Such interpretation has significant implications as population concentrates in cities and towns that lie near or on the Pedeapenninic Thrust Front, thus establishing whether this is the surface evidence of a

string of shallow thrust faults or rather of substantially deeper ramps implies significantly different ground shaking expected in future earthquakes. Continuous Quaternary activity is poorly expressed along the mountain front, yet it is witnessed by progressive deformation and tilting of river terraces and of outcropping syntectonic sediments (e.g. BOCCALETTI *et al*. 1985, 2011; PICOTTI and PAZZAGLIA 2008; WEGMANN and PAZZAGLIA 2009; PONZA *et al*. 2010; GUNDERSON *et al*. 2013, 2014).

In contrast, the buried outer thrust fronts of the Southern Alps exhibit simpler geometries that define a single wide arc running from Milano to the Garda Lake (Fig. 2). This outer arc is connected in the east to the N–NE-trending Giudicarie thrust system (CASTELLARIN and CANTELLI 2000; MASSIRONI *et al*. 2006), and is associated with high angle back-thrusts and small topographic highs emerging in the northern Po Plain that were first investigated by DESIO (1965) (see also LIVIO *et al*. 2009). Unlike the Northern Apennines mountain front, the foothills of the western Southern Alps do not comprise a tectonics-controlled mountain front, the only reported exception being a small area north of Milan (MICHETTI *et al*. 2012).

The active Northern Apennines thrust fronts exhibit a rather complex architecture, comprising potential seismogenic sources corresponding to thrust ramps lying at different structural levels. The occurrence of complex thrust geometries is due to the presence of several potential detachment levels in the sedimentary multilayer deformed by the thrust belt. Potential seismogenic sources include also a deep ramp located between about 12 and 22 km depth, well imaged in the CROP03 seismic reflection profile as a relatively high angle portion of a regional thrust (BARCHI *et al*. 1998; FINETTI *et al*. 2001). This thrust has been interpreted (1) as a major lithospheric structure, cutting the whole crust from the metamorphic basement up to the higher levels (BARCHI *et al*. 1998; LAVECCHIA *et al*. 2003, 2004); or (2) as one of the leading elements of the accretionary prism of the Apennine belt (BALLY *et al*. 1986; DOGLIONI *et al*. 1994). In the absence of geological evidence for shortening across this deep ramp, seismicity becomes the only available indicator of its activity. In fact, a number of earthquakes that occurred in the region northeast of the Northern Apennines crest at an
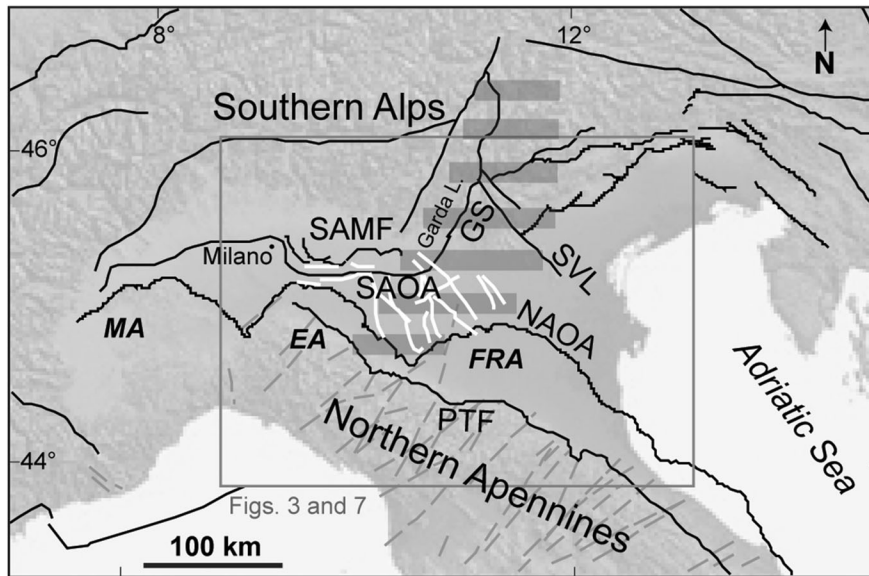
Figure 2
Structural sketch of the Po Plain. Key—*black lines* main structural elements (BIGI *et al*. 1992); *white lines* inherited faults (ROGLEDI, 2010); *dashed lines* transverse structures (e.g. CASTELLARIN *et al*. 1985); *grey horizontal shading* Trento Platform (CUFFARO *et al*. 2010). *SAMF* Southern Alps Mountain Front, *SAOA* Southern Alps Outer Arc, *GS* Southern Alpine Giudicarie system, *SVL* Schio-Vicenza Line, *NAOA* Northern Apennines Outer Arcs, *PTF* Pedeapenninic Thrust Front, *MA* Monferrato Arc, *EA* Emilia Arc, *FRA* Ferrara-Romagna Arc

estimated depth larger than ordinary (i.e. below 15 km) can be tentatively associated with this structure.

GPS studies show that convergence of Africa and Eurasia at the latitude of northern Italy is progressing at 3–8 mm/a (e.g. SERPELLONI *et al*. 2007) and that shortening across the Po basin proceeds at 1–3 mm/a, increasing from west to east (e.g. DEVOTI *et al*. 2011). Shortening is accomplished by counter-clockwise rotation of the Adriatic promontory, of which the Po basin is the northern termination, with respect to the European plate, with a pole of rotation located in the western Alps (ANDERSON and JACKSON 1987; WEBER *et al*. 2010 and references therein). Borehole breakout data show Sh-max axes perpendicular to the trend of the buried thrust fronts, in agreement with evidence from focal mechanisms (HEIDBACH *et al*. 2008; MONTONE *et al*. 2012; CARAFA and BARBA 2013).

2.2. Paleogeographic Control on Active Tectonics

The Cenozoic compressional architecture of both the Alps and the Apennines was strongly influenced by the inherited Mesozoic palaeogeographic setting

of its foreland, i.e. the Adriatic plate. This was characterized by carbonate platforms and deep marine basins, separated by roughly NNE-trending extensional faults developed during the Tethyan rifting of the ancient African passive margin (BOSELLINI and HSU 1973; BERTOTTI *et al*. 1993; CHANNELL 1996; RAVAGLIA *et al*. 2006; FANTONI and FRANCIOSI 2010). As a consequence of this inheritance, the lithosphere progressively deformed by the two thrust belts developed through stratigraphic sequences having uneven thickness and variable sedimentary facies (Table 1). Similarly, the basement rocks and the overlying sedimentary successions involved in the construction of the orogenic belts were characterized by strong lateral variations resulting in strong rheological contrasts (CUFFARO *et al*. 2010).

The variations in the properties of the subducting lithosphere controlled the evolution and the shape of the thrust fronts, that is to say, the development of salients and recesses, respectively, in front of basins and carbonate highs (MARIOTTI and DOGLIONI 2000; CUFFARO *et al*. 2010). In its turn, the linearity of the thrusts is related to the spacing between subsequent fronts, which is controlled by two important

Table 1

Summary of the characteristics of palaeogeographic domains in the central Po Plain, and control exerted on thrust geometry (observations mostly from CUFFARO *et al.* 2010, *reinterpreted*)

Characteristics	Basins (Lombardian/Belluno)	Platforms (Trento/Friuli)	Control on thrust geometry
Thickness of sedimentary sequence	Thicker	Thinner	–
Depositional facies	Deep marine	Shallow water carbonates	Number of detachment levels
Multiple detachment horizons	Yes	No	Presence of different wavelength structures
Dip of foreland monocline	Steep	Shallow	Spacing among thrust fronts
Depth of basal detachment	Deep	Shallow	Spacing among thrust fronts
Spacing among thrust fronts	Wide	Small	Development of salients and recesses
Trend/linearity of thrust front	Salients	Recesses	–

geometrical parameters of the orogen development: (1) the dip of the foreland monocline (MARIOTTI and DOGLIONI 2000; CUFFARO *et al.* 2010), and (2) the depth of the basal detachment, i.e. of the deepest thrust of the accretionary prism, usually lying at or near the transition between the crystalline basement and the sedimentary cover (BIGI *et al.* 2003). Along with the occurrence of shallower detachment horizons resulting from the vertical heterogeneities of the sedimentary sequence, these two properties of the subducting lithosphere have had a profound impact on the subsequent development of the thrust fronts on either side of the Po Plain. Table 1 summarizes how the development of the Alps and Apennines thrusts was controlled by the different characteristics of the sedimentary cover scraped off the foreland lithosphere, and thus of the accretionary prisms, and lists the resulting geometrical characteristics of the two thrust belts. In the following we will discuss some implications of this palaeogeographic control on active faulting.

The palaeogeographic inheritance may also have controlled (1) the reactivation in the present-day regime of *inherited* faults that bounded the Mesozoic rift system, and (2) the creation of foreland monocline panels characterized by different dip values; we speculate that such panels are separated by transfer zones characterized by strike-slip or transpressional motion and that they are usually perpendicular to the thrust fronts, or *transverse* (see Fig. 4 in CUFFARO *et al.* 2010; Fig. 2). Both types of faults, *inherited* and *transverse*, lie below the basal detachment of the accretionary prisms, being rooted in the lithosphere of the subducting foreland plate. In particular, we

expect to find the *transverse* faults only below the Apennines prism, because the initial momentum for their development and kinematics is induced by changes in the dip of the foreland monocline that are seen only beneath the Apennines (MARIOTTI and DOGLIONI 2000). Although the true geodynamic motivations definitely require further investigation, it is worth nothing that many of the transverse structures that cut across the Apennines (see Sect. 3.5) are geographically coincident with the zones where changes in the dip of the monocline are more pronounced. On this basis we propose a genetic relationship between them and the changes in dip of the foreland monocline.

Another class of structures rooted in the subducting Adriatic lithosphere includes a set of high-angle normal faults that have been recently discussed in works dealing with the reconstruction of the deep structural setting of the Po Plain (ROGLEDI 2010). These faults trend at a high angle with respect to the boundaries of the palaeogeographic domains (i.e., E–W to NW–SE); they have been mapped only in the "undeformed" portion of the Po Plain, i.e. beyond the outer fronts of the Alps and Apennines, but have been seen to extend below the Ferrara-Romagna Arc, possibly controlling its geometrical development (TOSCANI *et al.* 2013; BONINI *et al.* 2014). More specifically, TOSCANI *et al.* (2013) used analogue modeling to show that the location of the thrust ramps associated with the main anticlines has been controlled by the presence of steps produced by such high-angle faults in the basement morphology. These faults are seen to trend sub-parallel to the inferred direction of the hinge of the foreland monocline, and

may thus be optimally oriented for being reactivated due to the bending moment induced by the progressive increase of the monocline dip.

2.3. Seismicity and Earthquake Budgets

Any investigation of the seismicity of the Po Plain must carefully take into account that in the past both its natural and human landscapes were very different from today. Currently, the Po Plain is one of the most densely inhabited and cultivated areas of Italy; however, it once was covered by large wooded patches and marshlands, particularly south of the Po river. The reconstruction of the historical seismicity of the Po Plain is a complex task as it implies knowledge of the distribution of settlements and roads of the time. Few large earthquakes are known, the 1117 Veronese event being the only one with an inferred $M > 6.5$. Table 2 lists the main parameters of the largest historical earthquakes known to have occurred in the study region (MELONI et al. 1988; BOSCHI and GUIDOBONI 2003; GUIDOBONI et al. 2007; ROVIDA et al. 2011). In the following sections we will describe these events individually; each one is keyed to a code (IDx) and is tentatively assigned to a seismogenic source (Table 2).

Historical and instrumental catalogues (Tables 2, 3; CASTELLO et al. 2006; ISIDE WORKING GROUP 2010) show that the seismicity of the Po Plain concentrates along the foothills and the buried thrust fronts of the Northern Apennines and Southern Alps (Fig. 3), with a few notable exceptions including the 3 January 1117 earthquake (M_w 6.7, Veronese; ID 6). In general, the Northern Apennines fronts exhibit more historical earthquakes than the Southern Alps fronts. Seismicity is not evenly distributed, however, but increases from west to east, i.e. from the Monferrato to the Ferrara-Romagna Arc.

As recalled earlier, the present-day activity of the two mountain fronts is characterized by compressional focal mechanisms (http://www.bo.ingv.it/RCMT/searchRCMT.html; PONDRELLI et al. 2006), in full agreement with tectonic evidence. The 20 and 29 May 2012 Emilia earthquakes (M_L 5.9 and 5.8, respectively; Table 3), the largest earthquakes ever recorded instrumentally in the region, were caused by purely reverse slip with Sh-max axes perpendicular to

the buried thrust faults of the western Ferrara-Romagna Arc (e.g. SCOGNAMIGLIO et al. 2012, among others). The 2012 sequence was highly complex as a total of seven M 5+ earthquakes occurred during the long aftershock sequence (about 2,500 events), activating a ~ 50 km-long stretch of this buried outer front. Most of the earthquakes occurred between 1 and 12 km depth, above the basal detachment of the outer thrust fronts.

Earthquake sequences characterized by multiple, similarly large mainshocks are rather common over the whole Italian territory (see Table 4; Fig. 4). In addition to stress transfer among adjacent faults, their occurrence is likely related to the country's structural complexity, which causes the active fault systems to be broken up by numerous transverse structures; on the one hand these effectively limit the fault length that may rupture in a single large earthquake, but on the other hand they make the triggering of adjacent faults more likely. The complex sequences of the Po Plain and surrounding areas are shown in Tables 2 and 3 and are highlighted in Fig. 3b.

The Po Plain is also the locus of relatively deep earthquakes, conventionally assumed to coincide with all events located below 15 km depth. They occur below the basal detachment of the Northern Apennines and Southern Alps and, therefore, are not directly related to thrust tectonics. Two deep earthquakes that were felt over a large portion of northern and central Italy occurred in 2012 near Reggio nell'Emilia (25 January, M_L 5.0, depth 29.0 km; ID 41) and near Parma (27 January, M_L 5.2, depth 72.4 km; ID 63). Some historical earthquakes have affected large portions of the Po Plain and of the surrounding ranges, causing widespread damage and suggesting that deeper events are a common occurrence in the region. For example, the effects of the 13 January 1909, M_w 5.5, *Bassa Padana* earthquake (ID 16) are documented by 800 intensity points spread over an area of about 180,000 km^2, and by 297 points with intensity V (N_v) or larger (see panel b in Fig. 5). As first suggested by MELONI and MOLIN (1987), the 1909 event may have occurred at a relatively large depth, up to 40 km according to FACCIOLI (2013). These deep earthquakes may attain a significant magnitude, as shown by the 1909 event; in fact, their true magnitude is likely to be substantially larger than

Table 2

Parameters of the largest historical earthquakes ($M \geq 5.3$) that occurred in the study region (from Boschi *and* Guidoboni *2003;* Rovida *et al.* 2011*)*

ID	Date	Locality	Intensity	M_w	Multiple	Deep	Description/notes	Source(s)	Group	Tectonic structure
6	3 Jan 1117	Veronese	IX	6.7	N/A	N/A	Guidoboni and Comastri (2005) reported an independent earthquake (five intensity datapoints, I = VII–VIII) occurring on the same day, at the same hour, but 150 km to the south, in the territory of Pisa	ITIS140; ITCS076	#3	N/A
4	25 Dec 1222	Basso Bresciano	IX	5.8	N/A	N	The source of this earthquake could be either the thrust or the backthrust	ITCS002 or ITCS072 (see "notes")	#1	SAOA
33	30 Apr 1279	Rocca San Casciano	VII–VIII	5.6	N/A	N/A	Only five intensity datapoints. The CFTI catalogue reports 1 May 1279, M_w 5.9, I = VIII (at 2:00 a.m.). The CPTI does not report an event on 1 May, but rather two events on 30 April, one near Rocca San Casciano (I = VII–VIII), one near Camerino (I = X)	ITCS027	#2	DR
21	25 Jul 1365	Bologna	VII–VIII	5.4	N/A	N	Only one intensity datapoint available. Some sources refer to a 7 April shock, other more reliable sources to a 25 July shock. Two earthquakes struck Ferrara on 4 March (M_w 4.8) and Verona on 21 September (M_w 4.5)	ITCS047	#1	PTF
25	4 Aug 1383	Forlì	VII–VIII	5.4	N/A	N	Only one intensity datapoint available. Preceded by M_w 4.1 shock on 24 July 1383 near Parma	ITCS001	#1	PTF
1	26 Nov 1396	Monza	VII–VIII	5.4	N/A	N/A	Only one intensity datapoint available	ITCS002	#1	SAOA
27	3 Jul 1428	Predappio	VIII	5.5	N/A	N	–	ITCS001	#1	PTF
9	11 Jun 1438	Parmense	VIII	5.6	N/A	N	–	ITCS009	#1	EA
26	11 Aug 1483	Romagna meridionale	VIII–IX	5.7	N/A	N	Baratta (1901) reports a shock near Ferrara on 11 March 1483 and one near Cervia in 1484	ITCS001	#1	PTF
18	5 Jun 1501	Appennino modenese	IX	6.0	N/A	N	–	ITCS046	#1	PTF
20	3 Jan 1505	Bolognese	VIII	5.6	Y	N	The earthquake sequence started on 31 Dec 1504 (M_w 5.4) and ended 19 May 1505 (Boschi and Guidoboni 2003)	ITIS091; ITCS047	#1	PTF
13	17 Nov 1570	Ferrara	VIII	5.5	Y	N	The earthquake sequence started on 17 November 1570, went on with closely spaced aftershocks up to early 1572, and gradually died out until 17 March 1574, when a somewhat stronger shock caused further damage in Finale Emilia (Guidoboni et al. 2007)	ITIS090; ITCS050	#1	FRA
32	-- 1600	Palazzuolo	VII–VIII	5.4	N/A	N/A	Only one intensity datapoint available	ITIS141; ITCS012	N/A	N/A
14	19 Mar 1624	Argenta	VIII–IX	5.5	N/A	N	–		#1	FRA
34	22 Mar 1661	Appennino Romagnolo	X	6.1	N/A	N	–		4	AVML
22	11 Apr 1688	Romagna	IX	5.8	N/A	N	"… a large aftershock near Russi on 18 April 1688, a week after the mainshock" (Boschi et al. 1997). Not reported in the CFTI catalogue	ITIS100; ITCS109	#1	FRA
31	29 Oct 1725	Appennino Tosco-Emiliano	VIII	5.4	N/A	N	Located about 10 km to the east of the 14 September 2003, M_w 5.3, Appennino Bolognese earthquake (ID 66)		#4	PSL
23	4 Apr 1781	Romagna	IX–X	5.9	Y	N	–	ITIS093; ITCS001	#1	PTF
24	17 Jul 1781	Romagna	VIII	5.6	Y	N	–	ITCS001	#1	PTF

Table 2 *continued*

ID	Date	Locality	Intensity	M_w	Multiple	Deep	Description/notes	Source(s)	Group	Tectonic structure
38	25 Dec 1786	Riminese	VIII	5.6	N/A	N	The composite and individual sources responsible for this earthquake are described in KASTELIC et al. (2013)	ITIS035; ITCS039	#1	N/A
15	22 Oct 1796	Emilia orientale	VII	5.6	N/A	Y	This earthquake falls near the ITCS012, but at a deeper depth. The epicentral area is very close to that of the 1909 earthquake (ID 16)	–	#4	PSL
2	12 May 1802	Valle dell'Oglio	VIII	5.6	N/A	N	–	ITIS104; ITCS002	#1	SAOA
7	9 Oct 1828	Valle Staffora	VIII	5.8	Y	N	–	ITCS018	#1	EA
12	11 Sep 1831	Reggiano	VII–VIII	5.5	Y	N	The 1831 event fall at the junction between the EA and the FRA. Given the complex local seismotectonic pattern, this event may have been caused by two adjacent segments of the EA (ITCS009) or FRA (ITCS049)	ITCS009; ITCS049	#1	EA/FRA
11	13 Mar 1832	Reggiano	VII–VIII	5.5	Y	Y	The 1832 event is deeper than the ITCS009 and the ITCS049.	–	#4	EL
30	25 Jun 1869	Media Valle del Reno	VII–VIII	5.4	N/A	Y	This is a deep event that occurred near the junction of the Apennines deep ramps and the Viareggio-Val di Lima-Bologna transverse structure. As such, it might be equally well assigned to Groups #2 o #4	–	#2/4	DR/VVL
28	30 Oct 1870	Romagna	VIII	5.6	N/A	N		ITCS001	#1	PTF
29	17 Sep 1873	Liguria orientale	VI–VII	5.4	N/A	Y		–	#4	EL
35	17 Mar 1875	Romagna sud-orientale	VIII	5.9	N/A	N	The composite and individual sources responsible for this earthquake are described in KASTELIC et al. (2013)	ITCS036; ITCS039	#1	N/A
5	7 Jun 1891	Valle d'Illasi	IX	5.9	Y	N	Strong, damaging aftershocks reported on 11 Jun, 15 Jun, 29 Jun, 7 Aug, 21 Aug 1891	–	–	–
17	4 Mar 1898	Valle del Parma	VII–VIII	5.4	N/A	N	–	ITCS046	#1	PTF
3	30 Oct 1901	Salò	VIII	5.7	N/A	N	–	ITIS069; ITCS048	#1	SAMF
16	13 Jan 1909	Bassa Padana	VI–VII	5.5	N/A	Y	From MELONI and MOLIN (1987). This earthquake falls near the ITCS012, but the ITCS012 is not responsible for it, because the 1909 event is deeper than the ITCS012. Epicentral area very close to that of the 1796 earthquake (ID15)	–	#4	PSL
36	17 May 1916	Alto Adriatico	VIII	5.9	Y	N	The composite and individual sources responsible for this earthquake are described in KASTELIC et al. (2013)	ITIS034; ITCS030	#1	N/A
37	16 Aug 1916	Alto Adriatico	VIII	6.1	Y	N	The composite and individual sources responsible for this earthquake are described in KASTELIC et al. (2013)	ITIS033; ITCS030	#1	N/A
19	20 Apr 1929	Bolognese	VII–VIII	5.3	Y	N	BOSCHI and GUIDOBONI (2003)	ITIS103; ITCS047	#1	PTF
8	15 May 1951	Lodigiano	VI–VII	5.4	N/A	Y	From MELONI et al. (1988). This earthquake falls near the ITCS044, but the ITCS044 is not responsible for it, because the 1951 event is deeper than the ITCS044	–	#3	UDS
10	15 Jul 1971	Parmense	VIII	5.7	N/A	N	–	ITCS009	#1	EA

Each earthquake is tentatively assigned to a group and, when possible, also to a tectonic structure. Earthquakes belonging to the same sequence are listed together

Y yes, *N* no, *N/A* not applicable, *EA* Emilia Arc, *FRA* Ferrara-Romagna Arc, *PTF* Pedeapenninic Thrust Front, *SAMF* Southern Alps Mountain Front, *SAOA* Southern Alps Outer Arc, *DR* deep ramp, *SVL* Schio-Vicenza Line, *AVML* Arbia-Val Marecchia Line, *PSL* Prato-Sillaro Line, *VVL* Viareggio-Val di Lima-Bologna Line, *EL* Enza Line, *ML* Montone Line, *UDS* unknown deep structure

Table 3

Parameters of the largest instrumental earthquakes (M ≥ 4.5) in the study region (from ISIDe and Bollettino Sismico Italiano)

ID	Date	Time (hh mm ss)	Lat (°)	Lon (°)	Depth (km)	M_L	Multiple	Strike (°)	Rake (°)	Dip (°)	Focal mechanism reference	Notes	Source(s)	Group	Tectonic structure
62	09 Nov 1983	16 29 52	44.698	10.375	28.0[a]	4.8	N	14; 262	29; 129	43; 71	PON	Falls close to ITCS009, but at a deeper depth	–	#4	EL
44	24 Apr 1987	02 30 28	44.816	10.701	5.0	4.6	Y	–	–	–	–	–	ITCS049	#1	FRA
45	02 May 1987	20 43 54	44.811	10.723	24.0[a]	4.5	Y	45; 312	8; 163	73; 83	PON	Depth = 26 km according to PONDRELLI et al. (2006)	–	#4	EL
40	13 Sep 1989	21 54 01	45.821	11.170	40.4[a]	4.7	N	56; 146	0; −179	89; 90	PON	Depth = 26 km according to PONDRELLI et al. (2006)	–	#4	SVL
42	15 Oct 1996	09 56 01	44.798	10.817	5.0	4.9	N	217; 94	47; 132	53; 54	PON	Depth = 17.8 km according to PONDRELLI et al. (2006)	ITCS049	#1	FRA
67	10 May 2000	16 52 12	44.245	11.941	13.3	4.5	Y	309; 120	97; 84	43; 47	PON	Depth = 21 km according to PONDRELLI et al. (2002), 7–14 km according to CALDERONI et al. (2009)	ITCS001	#1	PTF
43	18 Jun 2000	07 42 07	44.808	10.707	5.0	4.5	N	308; 210	−161; −23	68; 73	PON	–	ITCS049	#1	FRA
66	14 Sep 2003	21 42 53	44.255	11.380	20.1°	5.0	N	246; 94	68; 107	40; 53	PON	°According to PICCININI et al. (2006). Depth = 23 km according to PONDRELLI et al. (2007). M_w 5.3 according to PICCININI et al. (2006)	ITCS058; ITCS027	#2	DR
39	24 Nov 2004	22 59 39	45.685	10.521	5.4[a]	5.2	N	57; 228	95; 83	56; 34	TDMT	[a]According to PESSINA et al. (2006). Depth = 15 km according to PONDRELLI et al. (2007)[a]	ITCS048	#1	SAMF
64	23 Dec 2008	15 24 22	44.544	10.345	22.9	5.2	N	93; 280	86; 95	50; 40	TDMT	–	ITIS135; ITCS027	#2	DR
65	23 Dec 2008	21 58 27	44.527	10.355	23.7	4.8	N	96; 292	80; 102	51; 40	TDMT	–	ITCS027	#2	DR
68	05 Apr 2009	20 20 53	44.230	11.913	24.5	4.6	N	121; 267	113; 64	50; 45	TDMT	Falls close to ITCS001, but at a deeper depth	–	#4	ML
59	17 Jul 2011	18 30 27	45.010	11.367	2.4	4.8	N	82; 269	85; 95	47; 43	TDMT	Foreshock of the 2012 Emilia sequence	ITCS050	#1	FRA
41	25 Jan 2012	08 06 37	44.871	10.510	29.0	5.0	N	76; 307	120; −46	63; 40	TDMT	–	–	#4	UDS
63	27 Jan 2012	14 53 13	44.523	10.009	72.4	5.2	N	286; 113	87; 97	63; 27	TDMT	Unrelated to thrust tectonics due to its depth	–	#4	UDS
54	20 May 2012	02 03 52	44.889	11.228	6.3	5.9	Y	103; 280	92; 88	46; 44	TDMT	2012 Emilia sequence	ITIS134; ITCS103	#1	FRA
55	20 May 2012	02 06 30	44.886	11.189	7.7	4.8	Y	–	–	–	–	2012 Emilia sequence	ITCS103	#1	FRA
56	20 May 2012	02 07 31	44.863	11.370	5.0	5.1	Y	–	–	–	–	2012 Emilia sequence	ITCS103	#1	FRA
46	20 May 2012	03 02 50	44.860	11.095	10.0	4.9	Y	279; 88	95; 81	64; 27	TDMT	2012 Emilia sequence	ITCS051	#1	FRA
57	20 May 2012	13 18 02	44.831	11.490	4.7	5.1	Y	291; 111	90; 90	49; 41	TDMT	2012 Emilia sequence	ITCS103	#1	FRA

Table 3 *continued*

ID	Date	Time (hh mm ss)	Lat (°)	Lon (°)	Depth (km)	M_L	Multiple	Strike (°)	Rake (°)	Dip (°)	Focal mechanism reference	Notes	Source(s)	Group	Tectonic structure
58	20 May 2012	17 37 14	44.876	11.382	3.2	4.5	Y	67; 237	96; 81	59; 31	TDMT	2012 Emilia sequence	ITCS103	#1	FRA
47	29 May 2012	07 00 03	44.851	11.086	10.2	5.8	Y	271; 81	93; 81	71; 20	TDMT	2012 Emilia sequence	ITIS107; ITCS051	#1	FRA
48	29 May 2012	08 25 51	44.901	10.943	3.2	4.5	Y	67; 274	76; 114	61; 32	TDMT	2012 Emilia sequence	ITCS051	#1	FRA
49	29 May 2012	08 27 23	44.854	11.106	10.0	4.7	Y	–	–	–	–	2012 Emilia sequence	ITCS051	#1	FRA
50	29 May 2012	10 55 57	44.888	11.008	6.8	5.3	Y	279; 97	92; 88	48; 42	TDMT	2012 Emilia sequence	ITCS051	#1	FRA
51	29 May 2012	11 00 02	44.873	10.950	11.0	4.9	Y	–	–	–	–	2012 Emilia sequence	ITCS051	#1	FRA
52	29 May 2012	11 00 25	44.879	10.947	5.4	5.2	Y	–	–	–	–	2012 Emilia sequence	ITCS051	#1	FRA
53	03 Jun 2012	19 20 43	44.899	10.943	9.2	5.1	Y	271; 81	93; 81	71; 20	TDMT	2012 Emilia sequence	ITCS051	#1	FRA
60	06 Jun 2012	04 08 31	44.434	12.354	25.6	4.5	N	110; 17	−129; −3	88; 39	TDMT	Falls near ITCS012, but at a deeper depth	–	#4	UDS
61	03 Oct 2012	14 41 28	44.750	9.631	32.2	4.5	N	308; 218	−175; −0	90; 85	TDMT	Falls near ITCS027, but at a deeper depth	–	#4	UDS

see Fig. 3 for earthquake locations. Focal Mechanism Reference: TDMT, http://cnt.rm.ingv.it/tdmt.html (from 24 Nov 2004); PON, PONDRELLI *et al.* (2006)

a Depth from CSI http://csi.rm.ingv.it/ (CASTELLO *et al.* 2006). Key to tectonic structures as in Table 2

the 5.5 estimate obtained from intensity data with automatic routine procedures, normally calibrated using upper crustal events. Therefore, the occurrence of such deep earthquakes cannot be ignored in developing a seismotectonic model of the Po Plain. Reasonably reliable depth data, however, are available only for earthquakes of the past 30 years. For older events we did not attempt a formal determination of the hypocentral depth, but we used the ratio of epicentral intensity and felt area and some expert judgment simply to separate deeper from shallower events. All earthquakes having M 5.3+ were thus carefully inspected and their intensity patterns compared (Fig. 5). Tables 2 and 3 list all earthquakes having an estimated depth below 15 km (highlighted in Fig. 3b).

Based on the data presented in Tables 2 and 3, we analyzed the seismic release of the Po Plain and of the Apennines foothills in terms of shallow vs. deep earthquakes, with reference to the historical and to the most recent seismicity (Table 5). Even considering the limited completeness of the historical record, the total observed moment release (3.31×10^{26} dyne cm, equivalent to a single M_w 7.0 earthquake) is rather small, especially if compared with the size of the investigated area (nearly 40,000 km^2). This result must be considered also in light of the known limited *earthquake efficiency* of the Po Plain as a whole (BARBA *et al.* 2010). Shallow earthquakes dominate the global seismic release, but deep events release more moment over the instrumental time window, in spite of the large contribution to shallow moment release supplied by the 2012 Emilia sequence. This suggests (a) that deep historical events are under-represented, or (b) that their magnitude is systematically underestimated, or a combination of these two end-members.

3. Seismogenic Sources of the Po Plain

3.1. Generalities

In this section we identify and characterize the main seismogenic sources that affect the Po Plain, i.e. those believed to be capable of producing $M \geq 5.5$ earthquakes (VALENSISE and PANTOSTI 2001b; BASILI *et al.* 2008).
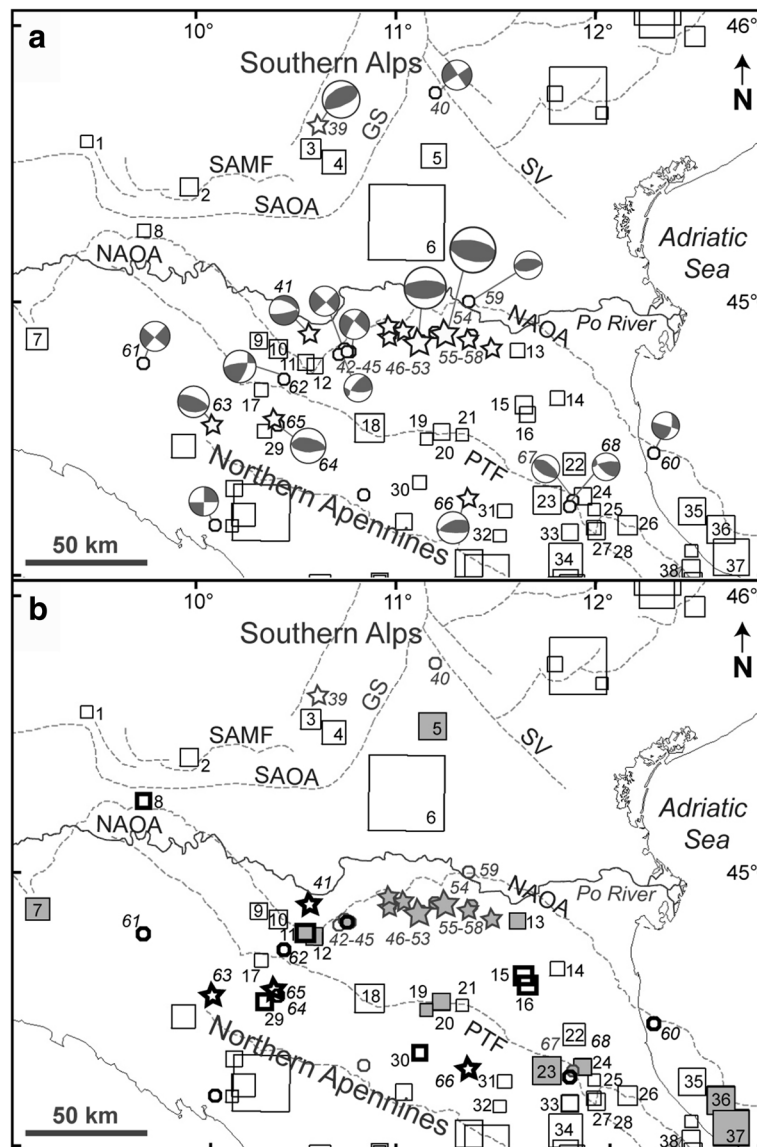
Figure 3

Historical and instrumental earthquakes from the CPTI11 catalogue (http://emidius.mi.ingv.it/CPTI11/), Bollettino Sismico Italiano (http://bollettinosismico.rm.ingv.it/) and ISIDe (http://iside.rm.ingv.it). Historical earthquakes have $M \geq 5.3$ and are shown with *squares*. Instrumental earthquakes have $M \geq 4.5$ and are shown with *circles* (M 4.5–4.9) or *stars* ($M \geq 5.0$). All earthquakes are keyed to Tables 2 and 3 through their ID. The IDs of instrumental earthquakes are shown in *italic*. **a** Available focal mechanisms (PONDRELLI *et al*. 2006; TDMT database: http://cnt.rm.ingv.it/tdmt.html; see Table 3 for focal parameters. **b** Earthquake locations: deep historical and instrumental earthquakes are shown in *bold*; multiple events are shown with a *grey* pattern

The approach employed for compiling the DISS database is based on substantial geological and geophysical evidence. The strategy is based on two categories of sources: the *individual seismogenic sources* (ISS; stored in the database with a unique code beginning with ITIS followed by a three digit number), and the *composite seismogenic sources* (CSS; stored in the database with a unique code beginning with ITCS followed by a three digit number). The first category includes individual fault segments that may be responsible for specific historical or instrumental earthquakes, whereas the second

Table 4

Summary of the 14 largest Italian earthquake sequences (M_w ≥ 5.7) of the past 50 years (1962–2012; data from CPTI11)

#	Date	Locality	M_w	Intensity	Kinematics	Multiple	Deep
1	1962, 21 August	Irpinia	6.1	IX	Normal	Y	N
2	1963, 19 July	Mar Ligure	6.0	VI	Thrust	Y	N
3	1968, 14 January	Valle del Belice	6.5	X	Thrust	Y	N
4	1976, 6 May	Friuli	6.4	IX–X	Thrust	Y	N
5	1978, 15 April	Golfo di Patti	6.0	VIII	Strike-slip	N	N
6	1979, 19 September	Valnerina	5.8	VIII–IX	Normal	N	N
7	1980, 23 November	Irpinia-Basilicata	6.7	X	Normal	Y	N
8	1984, 7 May	Appennino abruzzese	5.8	VIII	Normal	Y	N
9	1990, 5 May	Potentino	6.0	VII	Strike-slip	Y	Y
10	1997, 26 September	Appennino umbro-marchigiano	5.9	IX	Normal	Y	N
11	2002, 6 September	Palermo	5.8	VI	Thrust	N	N
12	2002, 31 October	Molise	5.8	VII–VIII	Strike-slip	Y	Y
13	2009, 6 April	L'Aquila	6.2	IX–X[b]	Normal	N	N
14	2012, 20 May	Emilia	5.9[a]	VIII[c]	Thrust	Y	N

See Fig. 4 for earthquake locations. Ten out of 14 of these sequences were multiple or complex events, i.e. they were characterized by at least two mainshocks whose magnitudes are within 0.3 units of each other, and two were "deep crustal". As each sequence included at least two— but also more—individual earthquakes above the given magnitude threshold, the total number of events considered is nearly 30

[a] M_L from ISIDE DATABASE (2010)

[b] TERTULLIANI *et al.* (2012a)

[c] TERTULLIANI *et al.* (2012b)

category represents fault systems that may include one or more ISSs. Detailed descriptions of the DISS sources and the multi-step procedure to identify their parameters can be found in BASILI *et al.* (2008, 2009), BURRATO *et al.* (2008), VANNOLI *et al.* (2012), KASTELIC *et al.* (2013) and on the database website (http://diss.rm.ingv.it/diss/UserManual.html).

Following the rationale of the DISS database, our work aims to represent an *updated* and hopefully complete portrait of the seismogenic sources recognized in the Po Plain. The completeness of the data is normally assessed with reference to the observed seismicity, making sure that no significant historical earthquake is left unassigned to a mapped source. The sparseness of the historical record, however, makes this assessment rather difficult and potentially misleading. In fact, the geological moment release that can be calculated for the Po Plain sources outpaces the observed historical moment release, suggesting that either the historical earthquake record is incomplete, or that their seismic coupling, or earthquake efficiency, is rather low (e.g. BARBA *et al.* 2010). This crucial aspect of the characterization of Po Plain sources is out of the scope of this paper, but will be the object of a future research effort.

As stated earlier, the seismogenic sources of the Po Plain originate from diverse geodynamic processes, occur at various depths, exhibit different kinematics, and, as such, they may induce very different ground shaking. For all of these reasons we subdivided them into four main groups (Fig. 6):

Group #1 shallow thrusts of the Alps and Apennines fold-and-thrust belts, which include fault systems occurring along the Alps and Apennines piedmont and more external fronts extending further into and beneath the Po Plain;
Group #2 deep ramps of the Apennines fold-and-thrust belts occurring beneath the Northern Apennines, northeast of the main drainage divide;
Group #3 inherited faults of the foreland, which includes older structures such as Mesozoic normal faults reactivated under the current compressional tectonic regime;
Group #4 transverse structures, faults that are roughly perpendicular to the main structural grain of the Northern Apennines thrust belt.

Sources of Group #1 pose a significant hazard as they occur at shallow depth and near the more densely populated areas (e.g. Figures 5, 6 in ZONNO
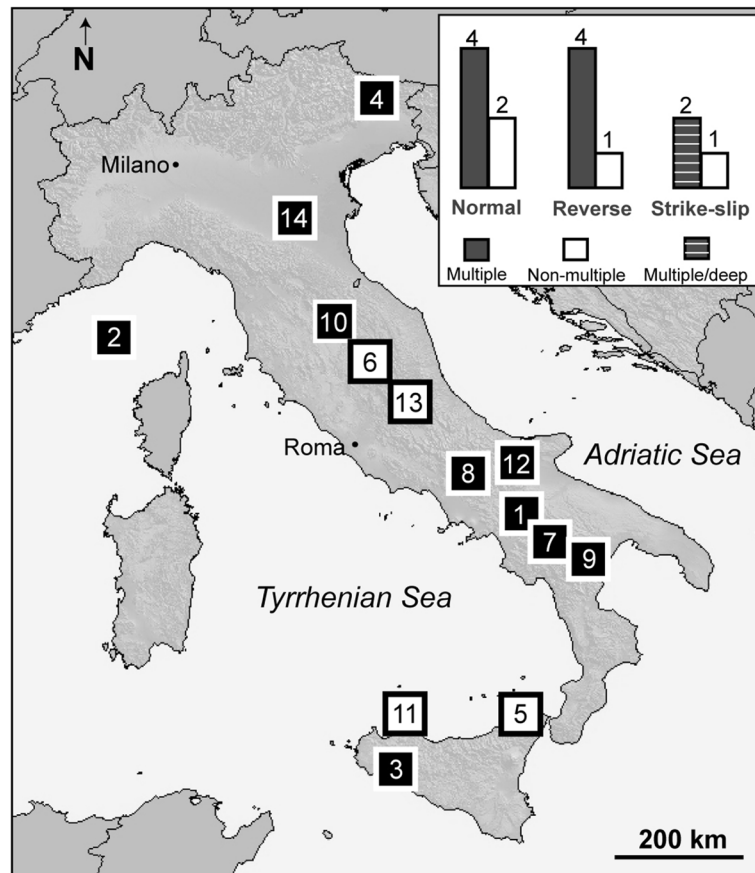
Figure 4
Largest Italian earthquake sequences ($M_w \geq 5.7$) of the past 50 years (1962–2012: data from CPTI11; see Table 4). A multiple sequence includes at least two earthquakes with comparable magnitudes. Multiple sequences are shown with *white lettering on black background*. *Inset* shows frequency of multiple events for different kinematics

et al. 2012). As such they are also somehow easier to identify despite their blind nature. Conversely, sources of Group #2 and #4 are usually responsible for moderate ground shaking over large areas, which partially compensates for the difficulties involved in locating them precisely. Finally, sources of Group #3 are perhaps the most dangerous as they are the most elusive, and thus their earthquake potential is most uncertain. As discussed in the following, the largest known earthquake of the Po Plain (3 January 1117, M_w 6.7) is believed to belong to this group. Although Groups #1 and #2 exhibit similar kinematics, we decided to keep them separate as their impact on ground shaking is very different due to their location and depth.

In this section we describe each CSS and ISS of the Po Plain included in the DISS database (version

3.2.0), from north to south and from west to east. For the study area, DISS lists 21 CSSs and 13 ISSs (Fig. 7). The sources in the DISS database belong to Groups #1, #2, and #3. Sources of Group #4 are still being evaluated for possible inclusion at a later time.

The following is an example of the data used and the reasoning adopted for characterizing an individual source, namely the source of the 4 April 1781, Romagna earthquake (M_w 5.9; ID 23, ITIS093). This upper crustal blind source (Group #1) is associated with gentle warping of the Apennines foothills and drainage anomalies. Its three-dimensional geometry was defined according to the following observations:

• the location was chosen by matching the larger faults that exist in the subsurface with the earthquake macroseismic pattern;
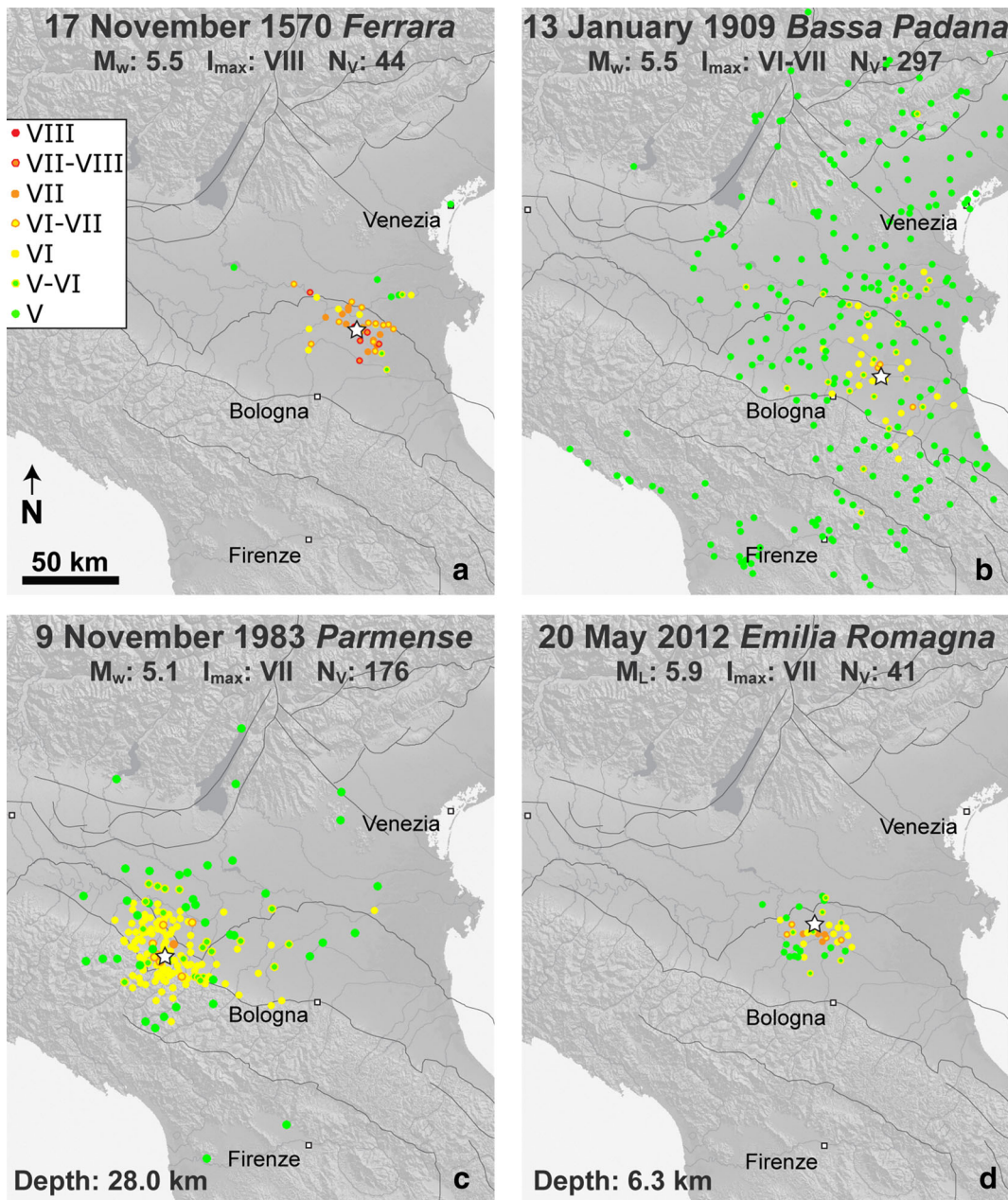
Figure 5

Maps of macroseismic intensity *V* and higher for the 1570 (**a**), 1909 (**b**), and 1983 (**c**) earthquakes (LOCATI *et al.* 2011), and for the 20 May 2012 (**d**) earthquake (TERTULLIANI *et al.* 2012b). The earthquake epicenters are shown with *white stars*. N_V is the number of the data points with intensity *V* or larger. Notice the large felt area of the 1909 and 1983 earthquakes, suggesting that the depth of these events is larger than the depth of the 1570 and 2012 events

- the strike was based on the local trend of the Pedeapenninic Thrust Front;
- the dip was set at 35° to the south, as suggested by subsurface data;

- the rake was assumed to be pure thrusting, based on geodynamic considerations constrained by mesostructures affecting recent deposits of the area;

Table 5

Summary of observed historical vs instrumental and shallow vs deep moment release in the Po Plain

Depth	Historical (1117–1982) dyne × cm	Instrumental (1983–2012) dyne × cm	Grand total dyne × cm
Shallow earthquakes (<15 km)	2.93×10^{26} (96 % of total historical)	2.14×10^{25} (84 % of total instrumental)	3.14×10^{26} (95 % of grand total)
Deep earthquakes (>15 km)	1.24×10^{25} (4 % of total historical)	3.96×10^{24} (16 % of total instrumental)	1.64×10^{25} (5 % of grand total)
Total moment release	3.05×10^{26} (92 % of grand total)	2.54×10^{25} (8 % of grand total)	3.31×10^{26} (100 % of total)

The earthquakes for which the depth could not be evaluated (Table 2) have been considered shallow events. See text for discussion
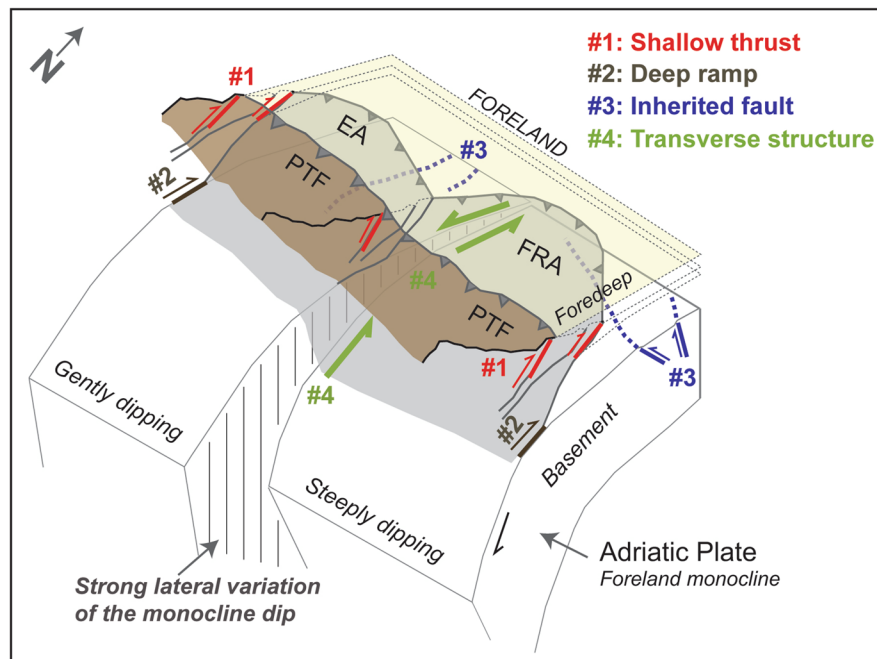
Figure 6
Sketch showing the four groups of seismogenic sources identified in this work. Group #1: shallow thrusts of the Northern Apennines, including the sub-emergent Pedeapenninic Thrust Front; Group #2: deep ramps of the Apennines thrusts; Group #3: inherited faults cutting the foreland; Group #4: transverse structures (or *lineaments*), interpreted as the response of the upper lithosphere to the variation of the dip of the regional monocline. *EA* Emilia Arc, *FRA* Ferrara-Romagna Arc, *PTF* Pedeapenninic Thrust Front

- the length was evaluated based on subsurface data and on the extent of landscape modifications induced by fault activity, and was constrained by scaling relationships with the magnitude of the 4 April 1781 earthquake;
- the down-dip width was assessed based on subsurface data and constrained using the magnitude of the 1781 earthquake;
- finally, the minimum and maximum depth were constrained by analysis of subsurface data.

Although the Po Plain has been the object of intense investigations since the inception of the DISS database, some potential sources of damaging earthquakes still need to be fully characterized. The existence of most of these new sources has been highlighted by the occurrence of earthquakes that, due to their location and hypocentral depth, cannot be associated with the Alps and Apennines thrust systems, and thus belong to Groups #3 and #4 (see above). Some of these sources have been included in
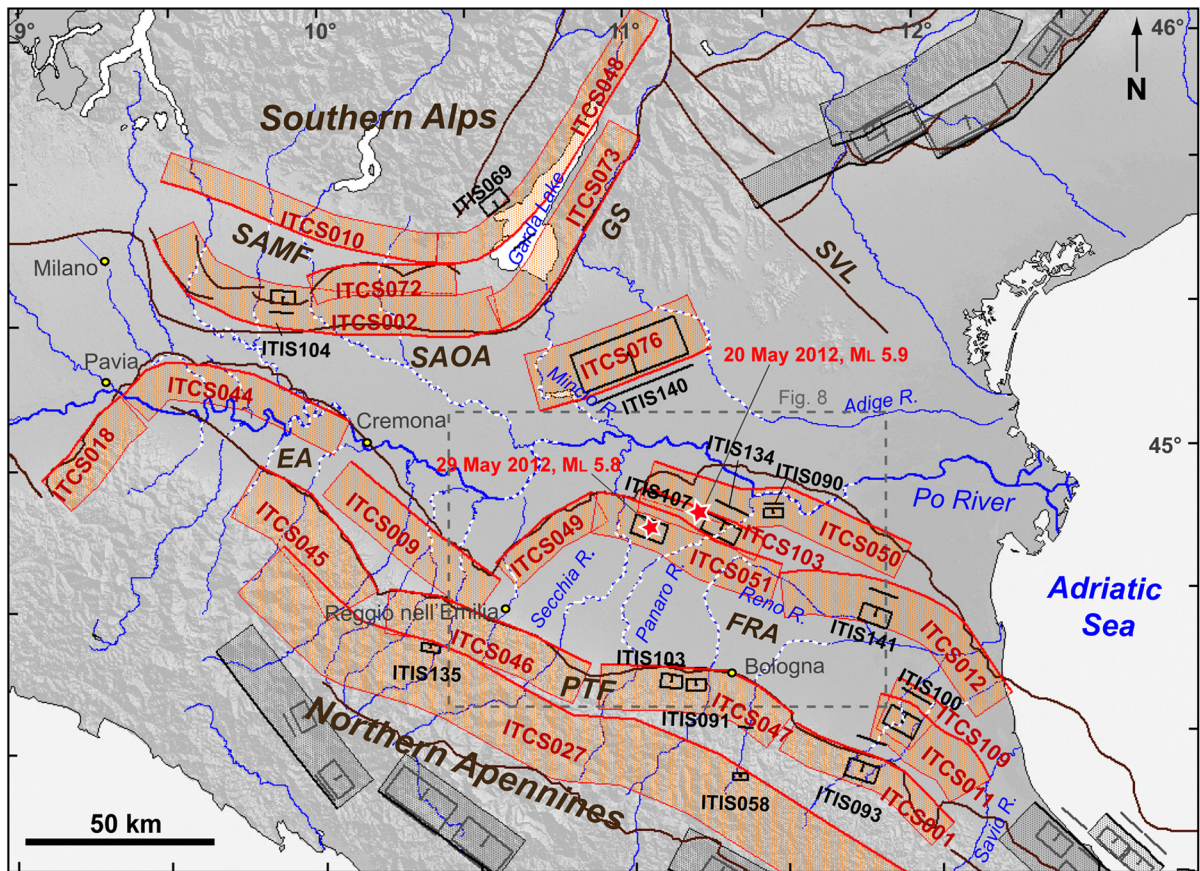
Figure 7

Composite and individual sources of the Po Plain, respectively shown as *red ribbons* and *black rectangles*. Reaches of anomalous drainage (from BURRATO *et al*. 2003) are marked in *white*. *SAMF* Southern Alps Mountain Front, *SAOA* Southern Alps Outer Arc, *GS* Southern Alpine Giudicarie System, *SVL* Schio-Vicenza Line, *PTF* Pedeapenninic Thrust Front, *EA* Emilia Arc, *FRA*: Ferrara-Romagna Arc

v. 3.2.0 of the DISS database, to be published in 2014, whereas others will be included in future releases.

3.2. *Shallow Sources of the Alps and Apennines Fold-and-Thrust Belts (Group #1)*

The seismogenic sources belonging to this group correspond to the shallow outer thrusts of the S-verging Central Southern Alps and of the N-verging Northern Apennines fronts that exist in the subsurface of the northern and southern Po Plain, respectively. Active compression in the Po Plain generates moderate seismicity along complex thrust systems and intervening strike-slip faults, an interpretation supported by available seismological, GPS, and stress field data. The occurrence of salients and

recesses along strike of the main thrust fronts creates boundaries to the dynamic earthquake propagation, causing segmentation of the compressional fault systems and, therefore, of the relevant CSSs.

The following description is based on a comprehensive instrumental catalogue (ISIDE WORKING GROUP 2010) and on historical catalogues (BOSCHI *et al*. 2000; GUIDOBONI *et al*. 2007; ROVIDA *et al*. 2011). For an in-depth description of each source and a complete list of references the reader may refer to the DISS website (http://diss.rm.ingv.it/diss/). All historical and instrumental earthquakes mentioned in the text are listed in Tables 2 and 3 and are highlighted in Fig. 3 through their identifier (ID).

Geodetic (D'AGOSTINO *et al*. 2008), seismological (ISIDE WORKING GROUP 2010; ROVIDA *et al*. 2011), and tectonic (MAESANO *et al*. submitted) evidence

suggest that convergence rates decrease from east to west across the whole Po Plain. Slow tectonic activity characterizes the Monferrato Arc (Fig. 2; e.g. MICH-ETTI et al. 2012 and references therein), in the westernmost portion of the Po Plain, and seismicity nearly disappears west of the longitude of Milan. For these reasons, the thrusts of the western end of the Southern Alps and of the Northern Apennines that fall in an area roughly corresponding to the administrative region of Piedmont have not yet been included in the DISS database. The only notable exceptions are the sources of the 2 and 16 April 1808, *Val Pellice* earthquakes (M_w 5.7, 5.6), located off the western end of the Monferrato Arc at the Alps foothills. Nevertheless, we believe that an in-depth analysis of the structures of the Monferrato Arc is mandatory as they show distinct evidence of mild recent activity and may thus generate very infrequent earthquakes.

3.2.1 Shallow Thrusts of the Central Southern Alps Margin

The southernmost compressional fault system of the Alps is a buried S-verging external arc that is clearly imaged by subsurface data between Milano and the Garda Lake (Fig. 7; PIERI and GROPPI 1981; CASSANO et al. 1986; FANTONI and FRANCIOSI 2010). The ITCS002 follows faithfully this poorly expressed blind thrust fault system. This portion of the Po Plain exhibits a lower seismic release compared to areas to the east and northeast (STUCCHI et al. 2008), yet we believe that structures belonging to the ITCS002 are responsible for the 12 May 1802 *Valle dell'Oglio* earthquake and possibly for the 26 November 1396 *Monza* earthquake (ID 2 and 1, respectively). The ITCS002 includes the ISS responsible for the 1802 earthquake (ITIS104: see Fig. 7; Tables 6, 7), whose location and geometry are based on geomorphological evidence supplied by BURRATO et al. (2003), on oil industry data (e.g. PIERI and GROPPI 1981; CASSANO et al. 1986; FANTONI and FRANCIOSI 2010) and on macroseismic data, that uniquely constrain the location of the thrust responsible for the 1802 event (ALBINI and ROVIDA 2010). Because of lack of suitable geological and macroseismic constraints, no ISS was derived for the 1396 event.

The S-verging front is accompanied by an active N-verging, high-angle back-thrust located southwest of the Garda Lake (Fig. 7; ITCS072). This source is associated with some of the small topographic highs first investigated and interpreted as culminations of anticlines related to buried compressional faults by DESIO (1965), later investigated by LIVIO et al. (2009) using shallow seismic reflection profiles. Both the frontal thrust (ITCS002) and its associated backthrust (ITCS072) are blind and their propagation toward the surface is expressed by secondary folds and fault planes (MICHETTI et al. 2012). The frontal thrust and the backthrust exhibit a cumulative displacement of 1,200 and 5,700 m, respectively. The cumulative slip rate dropped from 3.45 mm/a in the interval 1.6–1.2 Ma to 2.75 mm/a between 1.2 and 0.9 Ma and 0.47 mm/a from 0.9 Ma to present, with consistently larger deformation being accommodated on the backthrust with respect to the frontal thrust during the most recent period (LIVIO et al. 2009). Based on its macroseismic signature the 25 December 1222 *Basso Bresciano* earthquake (M_w 5.8; ID 4) may have been generated by either of these structures, both large enough to generate an event of that size, but the data do not allow for discrimination between these two scenarios.

The buried outer arc of the Southern Alps is connected to the east with the NNE-trending Giudicarie system (Figs. 2, 7), commonly interpreted as a regional transfer zone between the central and eastern Southern Alps (e.g. CASTELLARIN and CANTELLI 2000). The Giudicarie System developed along the border between Lombardian basin to the west, and the Trento platform to the east. The ITCS073 straddles the region east of the Garda Lake and corresponds to a SE-verging more external arc assumed to be an active ramp of the Giudicarie System. The role and geometry of this source is based on regional geological data (e.g. CASTALDINI and PANIZZA 1991). Historical and instrumental catalogues show scattered intermediate-size earthquakes across this region.

The ITCS048 is a S- to SE-verging internal thrust ramp of the Giudicarie System. The 30 October 1901 *Salò* earthquake occurred in this area (M_w 5.7; ID 3; PESSINA et al. 2013), followed about a century later and virtually in the same area by a further damaging

Table 6

Geometric and kinematic parameters of the ISSs

Source ID	Lat (°)	Lon (°)	Length (km)	Width (km)	Depth (km)	Strike (°)	Dip (°)	Rake (°)	Average slip (m)	Slip rate (mm/a)	Recurrence (a)	Associated earthquake	M_w	Group	Tectonic structure
ITIS069	45.634	10.514	7.0	5.0	6.5–9.0	231	30	90	0.35	0.10–0.50	700–3,500	30 Oct 1901	5.7	#1	GS
ITIS104	45.389	9.793	6.5	4.7	2.5–5.8	275	45	90	0.50	0.10–0.50	1,000–5,000	12 May 1802	5.7	#1	SAOA
ITIS140	45.271	10.992	30.0	14.0	3.0–12.0	248	40	90	1.00	0.10–0.50	2000–10000	03 Jan 1117	6.7	#3	–
ITIS090	44.890	11.498	5.1	4.0	1.4–4.5	88	50	90	0.35	0.10–0.50	700–3,500	17 Nov 1570	5.5	#1	NAOA-FRA
ITIS134	44.849	11.323	10.0	6.4	4.0–8.4	115	43	90	0.80	0.25–0.50	1,600–3,200	20 May 2012	6.1	#1	NAOA-FRA
ITIS107	44.848	11.074	9.0	5.9	4.0–7.0	108	30	90	0.40	0.50–1.04	385–800	29 May 2012	6.0	#1	NAOA-FRA
ITIS141	44.640	11.847	8.0	5.7	3.0–6.3	110	35	90	0.40	0.49–0.55	727–816	19 Mar1624	5.8	#1	NAOA-FRA
ITIS100	44.384	11.943	9.1	6.0	2.5–5.0	119	25	90	1.00	0.10–0.30	10,000–3,333	11 Apr 1688	5.8	#1	NAOA-FRA
ITIS103	44.478	11.158	5.6	4.3	2.0–4.5	98	35	90	0.40	0.10–0.50	800–4,000	20 Apr 1929	5.6	#1	PTF
ITIS091	44.470	11.243	5.0	3.9	2.0–4.2	90	35	90	0.40	0.10–0.50	800–4,000	03 Jan 1505	5.5	#1	PTF
ITIS093	44.264	11.807	8.4	5.7	4.5–7.8	108	35	90	0.40	0.10–0.50	800–4,000	04 Apr 1781	5.8	#1	PTF
ITIS135	44.547	10.334	4.9	3.9	18.5–21.7	98	55	79	0.30	0.10–0.50	600–3,000	23 Dec 2008	5.4	#2	DR
ITIS058	44.248	11.397	4.0	3.4	16.4–19.1	94	53	107	0.30	0.10–0.50	600–3,000	14 Sep 2003	5.3	#2	DR

The ID of each sources is the same used in the DISS database and shown in Fig. 7. Geometric coordinates refer to the centroid of the fault plane. Key to Groups: #1, shallow thrust; #2, deep ramp; #3, inherited fault. Key to tectonic structures

GS Southern Alpine Giudicarie system, *SAMF* Southern Alps Mountain Front, *SAOA* Southern Alps Outer Arc, *NAOA* Northern Apennines Outer Arc, *FRA* Ferrara-Romagna Arc, *PTF* Pedeapenninic Thrust Front, *DR* deep ramp

Table 7

Geometric and kinematic parameters of the CSSs

Source ID	Min depth (km)	Max depth (km)	Strike min (°)	Strike max (°)	Dip min (°)	Dip max (°)	Rake min (°)	Rake max (°)	Slip rate (mm/a)	M_w	Group	Tectonic structures
ITCS048	5	10	205	265	25	45	70	100	0.10–0.50	5.7	#1	GS
ITCS010	5	10	265	295	25	45	80	100	0.10–0.50	5.5	#1	SAMF
ITCS072	1	7	70	100	30	45	80	100	0.10–0.50	6.1	#1	SAOA backthrust
ITCS073	3	9	200	250	25	45	70	100	0.10–0.50	5.5	#1	GS
ITCS002	2	9	230	340	25	45	80	100	0.10–0.50	6.1	#1	SAOA
ITCS076	2	14	240	260	30	50	80	100	0.10–0.50	6.7	#3	–
ITCS050	1	8	85	115	25	55	80	100	0.10–0.50	5.5	#1	NAOA
ITCS018	2	8	30	50	20	45	60	90	0.10–0.54	5.5	#1	NAOA-EA
ITCS044	2	7	30	115	20	40	80	100	0.48–0.54[a]	5.5	#1	NAOA-EA
ITCS009	2	8	120	140	2	40	90	120	0.16–0.53[a]	5.6	#1	NAOA-EA
ITCS049	3	10	35	65	30	50	60	90	0.04–0.16[a]	5.5	#1	NAOA-FRA
ITCS103	3.5	10	110	120	40	50	80	100	0.25–0.50[a]	6.1	#1	NAOA-FRA
ITCS051	3	10	95	125	25	45	80	100	0.50–1.04[a]	6.0	#1	NAOA-FRA
ITCS012	2	8	85	150	20	40	80	120	0.49–0.55[a]	6.0	#1	NAOA-FRA
ITCS109	2.5	10	115	125	20	40	80	100	0.10–0.30[a]	6.0	#1	NAOA-FRA
ITCS011	2	10	110	140	25	45	80	100	0.15–0.33[a]	6.0	#1	NAOA-FRA
ITCS045	2	10	110	150	20	40	80	100	0.43–0.62[a]	5.5	#1	PTF
ITCS046	2	8	95	120	20	40	80	100	0.10–0.50	5.9	#1	PTF
ITCS047	2	8	85	135	20	40	80	100	0.10–0.50	6.0	#1	PTF
ITCS001	2	8	110	140	30	40	80	100	0.10–0.50	5.8	#1	PTF
ITCS027	12	22	90	160	20	55	70	110	0.10–0.50	6.2	#2	DR

Keys for groups and tectonic structures as in Table 6

[a] Slip rate from MAESANO et al. (submitted)

thrust faulting earthquake on 24 November 2004 (M_L 5.2, *Lago di Garda*; ID 39). The 1901 *Salò* earthquake has been associated with a segment of the ITCS048. The *Salò* Individual Source (ITIS069) is a 7 km-long, 231°-striking, NW-dipping blind thrust fault comprising a deeper ramp of the Giudicarie System. The depth of the ITIS069 was assigned based on felt reports of the 1901 earthquake. The 2004 earthquake occurred few kilometers north of the *Salò* source and was caused by slip on a portion of the ITCS048 source (PESSINA et al. 2006). The focal solution of the 2004 event further constrained and confirmed the parameters of the existing 1901 earthquake seismogenic source (ITIS069).

Farther to the west, the ITCS010 source is a S-verging, internal thrust front of the Southern Alps. Historical and instrumental catalogues show scattered moderate-size seismicity along strike of this thrust front. The role and geometry of this source are based on regional subsurface geological and geophysical data (e.g. CASTALDINI and PANIZZA 1991).

3.2.2 Shallow Thrusts of the Northern Apennines Margin

The northernmost compressional fault system of the Northern Apennines is arranged in a series of N-verging thrust fronts: from west to east, they are the outermost Emilia Arc, the Ferrara-Romagna Arc and, to the south, the Pedeapenninic Thrust Front (Fig. 7).

3.2.2.1 Emilia Arc The ITCS018 is the westernmost source of the Emilia Arc and lies south of Pavia, beneath the region extending from the Po River to the Northern Apennines foothills. Historical and instrumental catalogues show little seismicity in this area, with the notable exception of the 9 October 1828 earthquake (M_w 5.8; ID 7) that hit the Valle dello Staffora area close to the southwestern boundary of the source.

The ITCS044 source straddles the region between Pavia and Cremona. This area is characterized by

minor seismicity with the exception of the 28 July 1276 (M_w 5.1) earthquake. The San Colombano active fold, one of the few anticlines emerging in the Po Plain (TOSCANI *et al*. 2006), is the only surface expression of this source. Based on modeling of the geometry of the fold and of the associated growth strata, and adjusting the measured tectonic deformation with back-stripping of the differential compaction of the sediment, MAESANO *et al*. (submitted) calculated for this source an average slip rate of 0.48–0.54 mm/a over the past 1.81 Ma.

The ITCS009 source runs in a NW–SE direction from Cremona to Reggio nell'Emilia. A number of damaging earthquakes hit this area over historical times, the most important of which occurred on 11 June 1438 (M_w 5.6, *Parmense*; ID 9), 11 September 1831 (M_w 5.5, *Reggiano*; ID 12), and 15 July 1971 (M_w 5.7, *Parmense*; ID 10). Based on their peculiar intensity patterns, the 13 March 1832 (M_w 5.5, *Reggiano*; ID 11) and 9 November 1983 (M_L 4.8, *Parmense*; ID 62; Fig. 5c) earthquakes appear to be deeper than other events, and for this reason they may be tentatively interpreted as belonging to Group #4 (see below). The 1831 and 1832 earthquakes are close both in space and time, a phenomenon that is rather common for the Italian seismicity as shown also by the 2012 Emilia sequence (see Fig. 4; Tables 2, 3, 4). The average Pleistocene slip rate calculated for this source falls in the range 0.16–0.53 mm/a (MAESANO *et al*. submitted).

3.2.2.2 Ferrara-Romagna Arc   The ITCS049 is the westernmost source straddling the Ferrara-Romagna Arc. It extends to the NE of the city of Reggio nell'Emilia and follows the lateral ramp of the arc. This source accounts for a number of intermediate-size yet damaging earthquakes: 10 February 1547 (M_w 5.1, *Reggio Emilia*), 20 June 1671 (M_w 5.2, *Rubiera*), 12 February 1806 (M_w 5.2, *Novellara*), 25 Dec 1810 (M_w 5.3, *Novellara*), 24 April 1987 (M_L 4.6; ID 44), 15 October 1996 (M_w 5.4; ID 42), and 18 June 2000 (M_w 4.5, *Parmense*; ID 43). Based on the intensity pattern, the 2 May 1987 earthquake (M_w 4.5, *Reggiano*; ID 45) may have been generated by a deep source, and for this reason it may be assigned to Group #4 (see below). The average post-middle Pleistocene slip rate calculated for this source falls in

the range 0.04–0.16 mm/a (MAESANO *et al*. submitted).

The ITCS050 follows the most external thrust front of the Northern Apennines. Infrequent activity characterizes this portion of the plain with the only exception being the 22 February 1346 (M_w 4.9, *Ferrara*) and 17 November 1570 (M_w 5.5, *Ferrara*; ID 13; Fig. 5a) earthquakes. The 1346 event is crucial for the understanding of the earthquake potential of this region as it may have been located close to the events of the 2012 Emilia sequence; unfortunately, its macroseismic signature is insufficient to derive any firm conclusions concerning its focal parameters, starting with the magnitude, and further investigations are needed.

The most significant earthquake recorded in instrumental times occurred on 17 July 2011 (M_L 4.8; ID 59; Fig. 3; MASSA *et al*. 2012); it was located along the outermost front and had a reverse focal mechanism. In contrast with its low seismic activity, this area may hold a significant seismogenic potential, as suggested by subsurface geological and geophysical data (PIERI and GROPPI 1981; BOCCALETTI and MARTELLI 2004; MAESANO *et al*. submitted) and by evidence of active surface deformation (BURRATO *et al*. 2003). The 1570 Ferrara earthquake was associated with an individual thrust of this arc and identified as ITIS090 (DISS WORKING GROUP 2010). Its location and geometry are based on subsurface geological and geophysical information constrained by geomorphological observation of the occurrence of drainage anomalies on top of the buried structure.

The ITCS103 extends across the lower Secchia River and Panaro River valleys. Seismicity is sparse, but recent studies of historical sources suggest that a moderate-size earthquake, probably overlooked by previous workers, occurred on 6 April 1639 (CASTELLI *et al*. 2012). The 2012 Emilia sequence, which began with the 20 May 2012 earthquake (ITIS134 source), activated a 10 km-long portion of the ITCS103 (Fig. 7; Table 8). Subsurface data (App_Orien-1 seismic line, available from http://unmig. sviluppoeconomico.gov.it/videpi/) show well developed buried anticlines (Fig. 8), and the coseismic deformation of the ground-surface caused by the 20 May 2012 earthquake (BIGNAMI *et al*. 2012; PEZZO

Table 8

Geometric and kinematic parameters of the 20 and 29 May 2012 earthquakes as proposed by various investigators and institutions

References	Event	Length (km)	Width (km)	Min depth (km)	Strike (°)	Dip (°)	Rake (°)	Maximum slip (m)	Uniform slip (m)	M_o (Nm)	M_w	Method
GCMT	20 May	–	–	–	88/304	35/61	60/109	–	–	1.74E+18	6.1	CMT
GFZ	20 May	–	–	–	104/287	32/58	88/91	–	–	1.6E+18	6.1	MT
SLUEC	20 May	–	–	–	278/105	45/45	85/95	–	–	1.02E+18	5.94	RCMT
Pondrelli et al. (2012)	20 May	–	–	–	279/109	60/30	85/99	–	–	1.81E+18	6.11	QRCMT
Scognamiglio et al. (2012); TDMT	20 May	–	–	–	280/103	44/46	88/92	–	–	7.00E+17	5.86	TDMT
Malagnini et al. (2012)	20 May	–	–	–	285	45	90	–	–	–	5.63	MT
Saraò and Peruzza (2012)	20 May	–	–	–	282/97	51/39	93/86	–	–	1.37E+18	6.1	MT
Cesca et al. (2013)	20 May	–	–	–	103/286	25/65	87/91	–	–	1.89E+18	6.1	Seismological source inversion
Bignami et al. (2012)	20 May	19	4	4.0	115	43	90	–	0.80	–	6.0	Modeling of InSAR data
Serpelloni et al. (2012)	20 May	12.8	5.3	4.9	109	37.7	86	–	0.99	–	6.13	Modeling of GPS data
Pezzo et al. (2013)	20 May	34	11	1.0	114	40	90	1.20	–	–	–	Modelling of GPS and InSAR data
Tizzani et al. (2013)	20 May	14.8	3.6	4.3	114	40	80	–	0.70	1.12E+18	–	Modelling of GPS and InSAR data
GCMT	29 May	–	–	–	91/291	29/63	72/99	–	–	7.97E+17	5.9	CMT
GFZ	29 May	–	–	–	97/283	30/60	85/93	–	–	6.6E+17	5.8	MT
Pondrelli et al. (2012)	29 May	–	–	–	276/110	71/20	85/103	–	–	1.07E+18	5.96	QRCMT
Scognamiglio et al. (2012)	29 May	–	–	–	275/95	52/38	90/90	–	–	3.51E+17	5.66	TDMT
Malagnini et al. (2012)	29 May	–	–	–	270	45	85	–	–	–	5.44	MT
Saraò and Peruzza (2012)	29 May	–	–	–	274/97	64/26	89/93	–	–	7.66E+17	5.9	MT
Cesca et al. (2013)	29 May	–	–	–	103/281	27/63	92/89	–	–	8.81E+17	5.9	Seismological source inversion
Bignami et al. (2012)	29 May	16	6	3.1	102	20	95	–	0.34	–	5.9	Modeling of InSAR data
Serpelloni et al. (2012)	29 May	12.7	5.7	2.8	102	35.2	85.4	–	0.36	–	5.85	Modeling of GPS data
Pezzo et al. (2013)	29 May	32	7.5	1.0	95	45	85	0.54	–	–	–	Modelling of GPS and InSAR data
Tizzani et al. (2013)	29 May	20	2.6	4.5	97	40	90	–	0.80	1.24E+18	–	Modelling of GPS and InSAR data

Sources: *GCMT* global CMT catalogue, http://www.globalcmt.org/; *GFZ* GFZ moment tensor, http://geofon.gfz-potsdam.de/eqinfo/list.php; *SLUEC* Saint Luis University Earthquake Center, http://www.eas.slu.edu/eqc_mt/MECH.IT/; *QRCMT* quick regional moment tensors, http://autorcmt.bo.ingv.it/quicks.html; *TDMT* time domain moment tensor, http://cnt.rm.ingv.it/tdmt. html
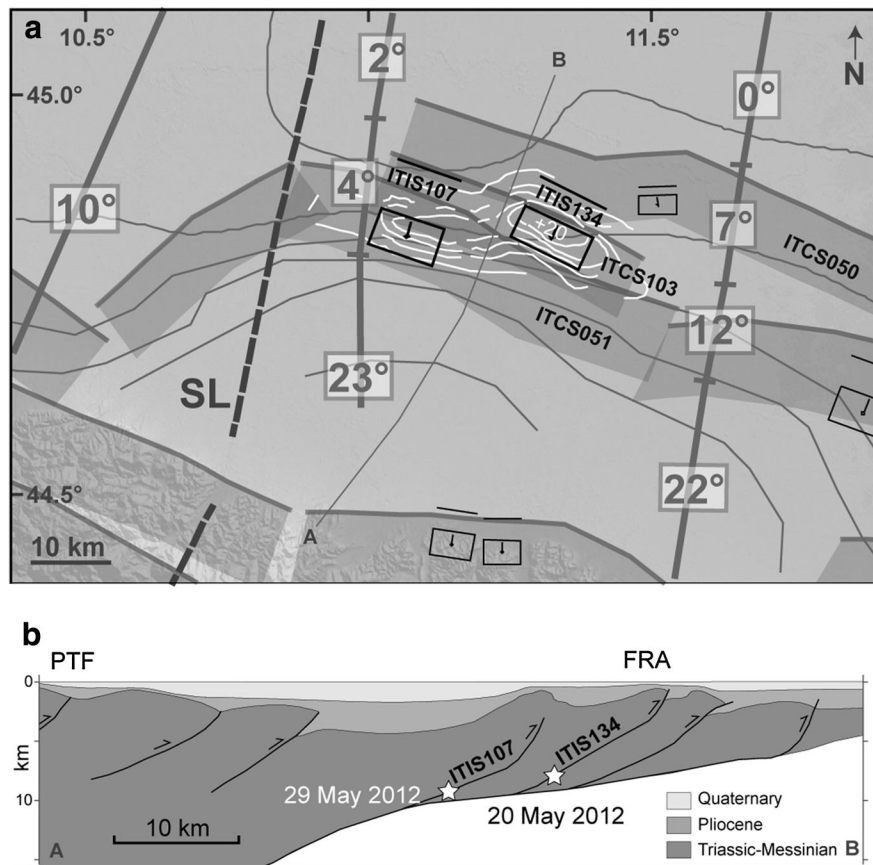
Figure 8

a Sketch of the seismogenic sources of the Northern Apennines thrust fronts in the epicentral area of the 2012 seismic sequence, and **b** simplified geological section. *Contour lines* of observed surface deformation from InSAR data (with maximum uplift value in cm) are shown in *white* (BIGNAMI *et al.* 2012). The isolines of the dip of the foreland monocline are shown in *grey* (MARIOTTI and DOGLIONI 2000, reinterpreted): *numbers* indicate the dip of the Apennines monocline along the three sections (MARIOTTI and DOGLIONI 2000). The *dashed line* shows the Secchia Line (SL; northern segment: CASTALDINI *et al.* 1979; southern segment: CASTELLARIN *et al.* 1985)

et al. 2013, TIZZANI *et al*. 2013) is consistent with the characteristics of this seismogenic source.

The ITCS051 source encompasses the region across the lower Secchia River and Reno River valleys. CASTELLI *et al*. (2012) found historical evidence of moderate-size earthquakes that occurred in the area on 15 December 1761 and 11 May 1778. Subsurface data (e.g. CASSANO *et al*. 1986) show a number of well developed buried folds, the largest being the Mirandola anticline. BURRATO *et al.* (2003) proposed that a segment of the blind thrust imaged by subsurface data could be the source of infrequent M_w 5.5 or larger earthquakes in this area. The ITIS107 *Mirandola* source was identified on the basis of tectonic geomorphology observations and considerations, stressing the relationships between peculiar

drainage anomalies and a major buried active fault. Between Sorbara and San Prospero, the Secchia and Panaro rivers are drawn toward the earthquake epicentral area, whereas between San Felice sul Panaro and Mirandola they are pushed away from it. The existence of the ITIS107 source explains well the drainage (1) being attracted towards a locally subsiding area, and (2) being diverted away from an uplifted area. As the ITIS107 *Mirandola* was not associated with any historical earthquake; its assigned M_w 6.2 was derived only from the geologically constraints on the source size. This potential seismogenic source was included in the version of DISS database presented in 2000 (v. 1.0) and published in 2001 (v. 2.0, VALENSISE and PANTOSTI 2001b). The 29 May 2012 earthquake and its associated sequence

were caused by the activation of a portion of ITCS051. Notably, the 29 May mainshock may have been caused by the ITIS107 itself, and the ITCS051 source has been identified in DISS as the source responsible for the uplift of the Mirandola anticline. The coseismic deformation of the ground-surface caused by the 29 May 2012 earthquake and measured by SAR interferometry (see BIGNAMI et al. 2012, among others; Table 8) is consistent with predictions based on modeling of the characteristics of this seismogenic source. The source responsible for the 29 May 2012 event is described in detail in BURRATO et al. (2012). The average middle Pleistocene slip rate calculated for this source falls in the range 0.50–1.04 mm/a (MAESANO et al. submitted); this figure is specifically based on observations carried out along sections across the buried Mirandola anticline.

The ITCS012 source extends across the lower Reno River valley and strikes E–W to NW–SE. A number of moderate-size to damaging events have affected the area, including the 19 March 1624 (M_w 5.5, *Argenta*; ID 14) and 30 December 1967 (M_w 5.2, *Bassa Padana*) earthquakes. The 1624 earthquake is assumed to have been caused by the ITIS141 source. The ITCS012 corresponds to the most advanced of the Romagna thrust fronts and marks their termination to the east. The average middle Pleistocene slip rate calculated for this source falls in the range 0.49–0.55 mm/a (MAESANO et al. submitted).

The ITCS109 and ITCS011 extend over the portion of the Po Plain located east of Bologna. These fault systems correspond to NE-verging inner arcs located north of the Romagna foothills. Local seismicity is sparse and moderate in size, with the exceptions of the 10 July 1591 (M_w 5.2, *Forlì*) and 11 April 1688 earthquakes (M_w 5.8, *Romagna*; ID 22). Subsurface data (PIERI and GROPPI 1981; CASSANO et al. 1986; FANTONI and FRANCIOSI 2010) and evidence from a number of investigators point to fast active surface deformation resulting from the tectonic activity of the buried anticlines at the rear of the Ferrara-Romagna Arc. The average middle Pleistocene slip rate calculated for the ITCS011 falls in the range 0.15–0.33 mm/a (MAESANO et al. submitted). A segment of the ITCS109 (ITIS100) has been hypotesized as the source of the 11 April 1688,

Romagna earthquake. The location and geometry of this source were based on a comparison between the 1688 damage pattern and the location of major thrusts of the Ferrara-Romagna Arc in the subsurface, whereas its size was scaled to the M_w 5.8 obtained from intensity data.

3.2.2.3 Pedeapenninic Thrust Front The ITCS045, ITCS046, ITCS047, and ITCS001 sources run along the Pedeapenninic Thrust Front. They are NE-verging thrusts and folds generating the geomorphology of the Emilia-Romagna foothills (e.g. PONZA et al. 2010). Their present-day activity and seismogenic potential are suggested by subsurface geophysical data (PIERI and GROPPI 1981; CASSANO et al. 1986; BRUNO et al. 2011), evidence of drainage anomalies (AMOROSI et al. 1996), structural data (CASTELLARIN et al. 1985), morphotectonic analyses (PICOTTI and PAZZAGLIA 2008; PONZA et al. 2010; GUNDERSON et al. 2013), and evidence from past seismicity.

In our interpretation, the ITCS045 source marks the western end of the active Northern Apennines mountain front and has been the locus of few moderate-size earthquakes. The ITCS046 source extends across the region south of the city of Reggio nell'Emilia. Moderate-size and damaging earthquakes are numerous but sparse all across the area. Key earthquakes have occurred on 5 June 1501 (M_w 6.0, *Appennino Modenese*; ID 18), 9 December 1818 (M_w 5.2, *Langhirano*), and 4 March 1898 (M_w 5.4, *Valle del Parma*; ID 17). A moderate-size earthquake took place in 91 B.C. (M_w 5.7, *Modena-Reggio Emilia*) just north of this area, but its location is inevitably very uncertain (GUIDOBONI et al. 2007).

The ITCS047 extends across the upper Reno River valley, near the city of Bologna. The region is the locus of both moderate-size and damaging earthquake, mostly concentrated in its central and western portions. Key earthquakes occurred on 25 July 1365 (M_w 5.4, *Bologna*; ID 21), 20 July 1399 (M_w 5.1, *Modenese*), 3 January 1505 (M_w 5.6, *Bolognese*; ID 20), and 20 April 1929 (M_w 5.3, *Bolognese*; ID 19). BOSCHI and GIUDOBONI (2003) listed a number of moderate-size earthquakes that occurred near Bologna and its surroundings, including the 4 May 1433 (I = VI MCS), 6 February 1455 (I = VIII MCS), and 20 December 1455 (I = VI–VII

MCS) events. The 1929 and 1505 earthquakes have been associated with two portions of ITCS047 (ITIS103 and ITIS091, respectively). The ITIS103 and the ITIS091 sources are consistent with evidence of recent activity of the Pedeapenninic Thrust Front southwest of Bologna, which includes triangular facets, the presence of a growing anticline and a fault scarp across Holocene alluvial fans. The rivers flowing southeast of Bologna are characterized by erosional behavior and uplifted terraces (e.g. PICOTTI and PAZZAGLIA 2008). The location of the ITIS103 source is consistent with the largest damage reported following the 1929 earthquake. In the same area, GUIDOBONI et al. (2007) reported landslides, surface fractures, and enhanced gas discharge in nearby gas and mud volcanoes. The location of the source is constrained to the east by the proximity to the ITIS091 source, assumed to be the causative fault of the 3 January 1505 earthquake located between Zola Predosa and Bologna. Also, this source was set in agreement with the largest reported damage, and also for this area GUIDOBONI et al. (2007) reported diffuse landslides and surface fractures, while PRESTININZI and ROMEO (2000) reported liquefaction at Zola Predosa. The location of the ITIS091 source is constrained to the east by the location of another M_w 5.1 earthquake on 20 January 1505 (BOSCHI and GUIDOBONI 2003), and to the west by the location of the 20 April 1929 earthquake.

The ITCS001 source encompasses the region to the west of the upper Savio River valley. The region exhibits a concentration of both moderate-size and damaging earthquakes, including the 4 August 1383 (M_w 5.4, *Forlì*; ID 25), 3 July 1428 (M_w 5.5, *Predappio*; ID 27), 11 August 1483 (M_w 5.7, *Romagna meridionale*; ID 26), 4 April 1781 (M_w 5.9, *Romagna*; ID 23), 17 July 1781 (M_w 5.6, *Romagna*; ID 24), 21 September 1813 (M_w 5.3, *Romagna centrale*), 30 October 1870 (M_w 5.6, *Romagna*; ID 28), and 19 February 1911 (M_w 5.2, *Romagna meridionale*) events. The two 1781 earthquakes comprise an additional case of "twin events" discussed earlier on (see Fig. 4; Table 4). An earthquake swarm caused by slip on WNW–ESE striking blind thrusts between 7 and 14 km depth occurred in the area in 2000 (mainshock on 10 May 2000, M_L 4.5; ID 67). Based on observations of this sequence CALDERONI et al. (2009) provided new insight

on the role of fluids involved in the seismogenic process. A segment of the ITCS001 is believed to be responsible for the 4 April 1781 earthquake (ITIS093). Our source model is consistent with the location of the largest damage, and the macroseismic pattern is suggestive of a shallow fault (see the Sect. 3.1 for further details on this source). In contrast, PICOTTI et al. (2009) hypothesized that the 1781 earthquake was produced by the *Riolo Terme* normal fault. This fault belongs to a set of normal faults occurring along the foothills of the Northern Apennines, showing evidence of recent activity and interpreted to develop along the outer shell of a large, growing anticline as a consequence of a local inversion of the stress field (PICOTTI et al. 2009). This interpretation, however, does not agree with the damage distribution of the 4 April 1781 earthquake as it leaves most of the largest intensity data points in the footwall of the fault, in contrast with the common observation that the strongest shaking due to a dip-slip fault is experienced in the hangingwall (ABRAHAMSON and SOMERVILLE 1996). The seismogenic potential of the normal faults investigated by PICOTTI et al. (2009) is also at odds with their secondary nature, as they most likely formed as a consequence of surface warping driven by a blind thrust. This further stresses the hypothesis that the main seismogenic source is the blind thrust itself.

3.3. Deep Ramps of the Northern Apennines (Group #2)

The ITCS027 source is the most internal compressional composite source of the Northern Apennines. As anticipated in the Sect. 2.1, this source corresponds with a deep thrust that has been interpreted either as a major lithospheric structure cutting the whole crust from the metamorphic basement up to the higher levels, or as one of the leading elements of the accretionary prism of the Apennines belt. In the latter case the earthquakes would generate at the interface where the sedimentary cover is delaminated from the metamorphic basement. The main differences between the two end-members in the interpretation of the structural style of this region concern the position of the basal decollement level of the thrusting and the presence or absence of secondary decollement levels. Some

investigators (e.g. BALLY *et al.* 1986; DOGLIONI *et al.* 1994) indicate that the bottom of the Triassic evaporites as the basal decollement level, and propose that the Palaeogene–Neogene evolution of the belt has been characterized by thin-skinned tectonics. Others (e.g. BARCHI *et al.* 1998; FINETTI *et al.* 2001; LAVECCHIA *et al.* 2003) contend that the upper portion of the magnetic basement was also involved in thrusting, and support a thick-skinned structural style. In general, most workers (e.g. BARCHI *et al.* 1998; COWARD *et al.* 1999) agree on the presence of two main decollement levels inside the accretionary prism; the lower coincides with the bottom of the Triassic evaporites, on which the major thrust ramps originate, whereas the shallower coincides with the Marne a Fucoidi or Scaglia Cinerea formation, the level marking the detachment of minor thrusts sheets. As a result, the deformation style is also characterized by two main structural levels: (1) a lower level, consisting of deep and broad anticlines produced by deep thrust faults detaching on the Triassic evaporites and causing deformation of the overlying Mesozoic-Palaeogene carbonate succession evaporates; and (2) an upper level, made up of small and tight anticlines lying in the axial zones of the major anticlines and developed in the Neogene foredeep sediments. These latter structures exhibit a more complex structural pattern with respect to the deeper ones.

We believe that the northwesternmost portion of the ITCS027 source is responsible for the 25 June 1869 (M_w 5.4, *Media Valle del Reno*; ID 30), 14 September 2003 (M_w 5.3, *Appennino bolognese*; ID 66) and 23 December 2008 (M_L 4.8 and 5.4, *Parmense*; ID 64 and 65) earthquakes; the two most recent events exhibit a depth around 20 km and reverse focal mechanisms. The ITCS027 contains two ISSs, the ITIS135 and ITIS058 (Fig. 7). The ITIS135 is a 4.9 km-long, N98°-striking, S-dipping blind thrust fault that is held responsible for the 23 December 2008, M_w 5.4 earthquake. Its strike, dip and rake were chosen based on the earthquake focal solution (http://www.bo.ingv.it/RCMT/searchRCMT.html; PONDRELLI *et al.* 2011). The mainshock occurred at 24 km depth and most of the aftershocks occurred below 20 km depth (ISIDE WORKING GROUP 2010). As a result, the earthquake was felt over a large area (INGV, http://www.

haisentitoilterremoto.it/repository/2205007630/index. html; ERCOLANI *et al.* 2009).

The ITIS058 source is a 4 km-long, N94°-striking, S-dipping blind thrust fault thought to be the source of the 14 September 2003, M_w 5.3, earthquake. Similarly to the previous event, this deeper earthquake was felt over a large area, but caused no damage, and its focal parameters were chosen based on its RCMT focal solution. The mainshock was located at 20 km depth; the aftershocks occurred between 15 and 22 km depth and showed a sharp NE–SW alignment. Almost all of the aftershocks were located E of the mainshock. Based on the depth distribution of the aftershocks, PICCININI *et al.* (2006) suggested that the mainshock occurred on a NW-dipping back-thrust of the Northern Apennines. Conversely, we propose that the mainshock occurred on a SE-dipping plane, but then triggered an aftershock sequence along a fault probably corresponding with the Prato-Sillaro Line (PSL in Fig. 9; see Sect. 3.5 for further details).

3.4. Inherited Faults of the Foreland (Group #3)

Inherited faults belonging to the Mesozoic rift system and reactivated in the present-day tectonic regime represent the potential sources of perhaps the largest earthquakes that occur beneath the Po Plain, such as the 3 January 1117, M_w 6.7 *Veronese* earthquake (ID 6; Table 2; Fig. 3) located to the south of the Lessini mountains. Updated macroseismic intensity data for this event clearly show that the area of greatest damage lies in the Adige plain south of Verona (GUIDOBONI *et al.* 2005). Perhaps because of the widespread effects and the fact that the epicentral area has always been considered an undeformed portion of the foreland, all previous investigators located the 1117 event 30–50 km away from the macroseismic epicenter; either along exposed fault systems, such as the Veneto thrust front (GALADINI *et al.* 2005), or along known buried faults, such as the thrust underlying the Piadena anticline (GALLI 2005).

Our hypothesis for the source of the 1117 earthquake is included in DISS as ITIS140 (ITCS076; Fig. 7). Its identification and parameters are based on the analysis and joint interpretation of subsurface
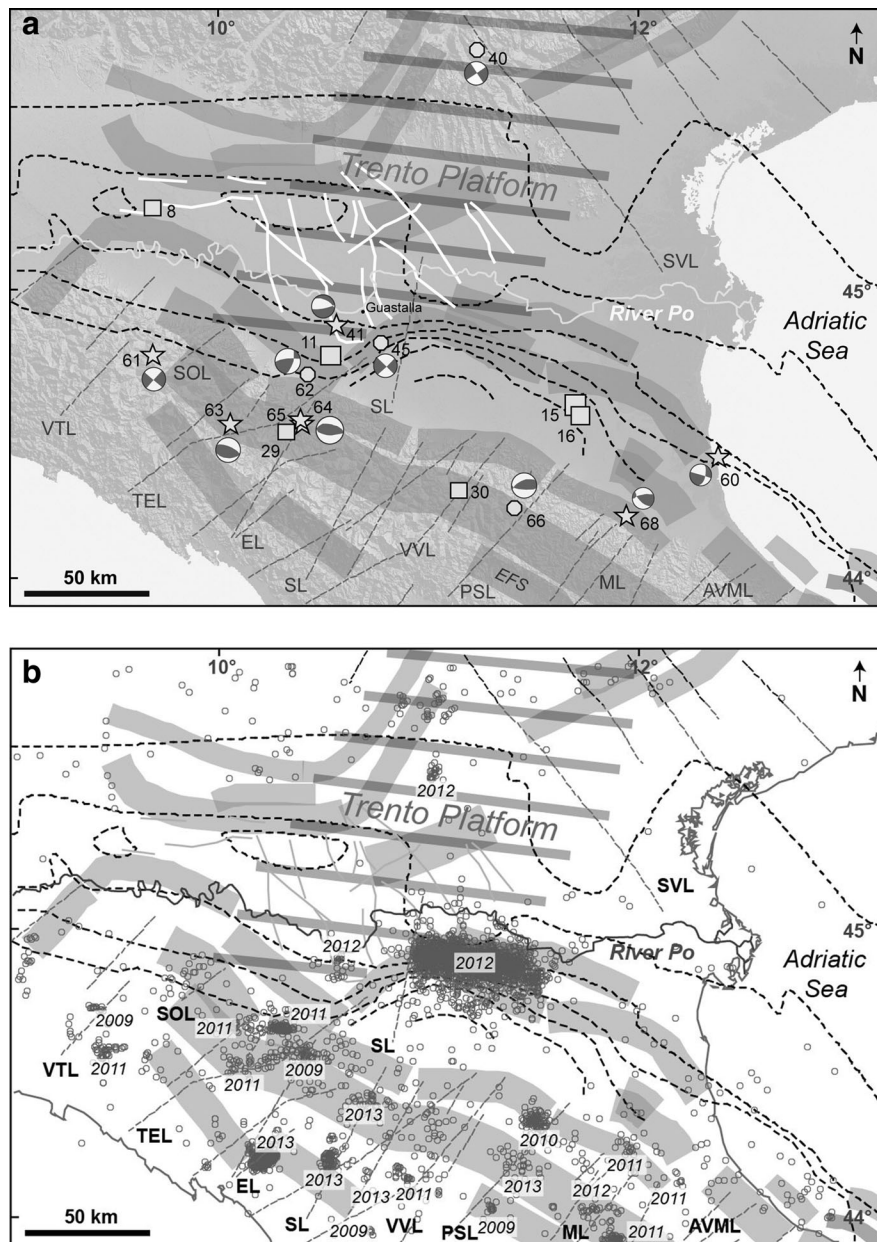
Figure 9

Seismotectonic sketch of the Po Plain and principal paleogeographic elements. The isolines contour the dip of the foreland monocline (MARIOTTI and DOGLIONI 2000, reworked). *Grey ribbons* CSSs (DISS WORKING GROUP 2010). *White* (in **a**)/*grey* (in **b**) *solid lines* inherited faults from ROGLEDI (2010). *Grey shaded area* Trento Platform from CUFFARO *et al.* (2010). *Grey dashed lines* transverse structures; *SVL* Schio-Vicenza Line, *VTL* Val Trebbia Line, *SOL* Stirone-Ongina Line, *TEL* Taro-Enza Line, *EL* Enza Line, *SL* Secchia Line, *VVL* Viareggio-Val di Lima-Bologna Line, *PSL* Prato-Sillaro Line, *ML* Montone Line, *AVML* Arbia-Val Marecchia Line (from CASTELLARIN *et al.* 1985). **a** Map showing deep earthquakes with *M* > 4.5. *Squares* historical (Table 2). *Circles* and *stars* instrumental (Table 3) with *M* > 4.5 and depth >15 km (*circles M* 4.5–4.9; *stars M* ≥ 5.0). *EFS* Etrurian Fault System. **b** Map of earthquake sequences (*M* ≥ 2.0) that occurred in the interval 1 January 2009–31 December 2013 (ISIDE WORKING GROUP 2010)

data and morphometric indicators of the drainage pattern in the portion of the Adige plain where the largest damage was reported. In the framework of an

investigation of drainage patterns in the entire Po Plain, BURRATO *et al.* (2003) found that the Mincio and Adige river courses exhibit an anomalous

behavior. A subsequent study identified drainage and topographic anomalies also along minor rivers of the area, all compatible with the existence of a buried reverse fault which would induce vertical relative ground motions in the region lying between the mid-course of the Adige and Mincio rivers. The best fitting reverse fault derived from modeling of these anomalies was seen to match a buried structure recognized in seismic exploration lines, a former extensional fault that is suitably oriented (ENE–WSW) for being reactivated under the current stress field. For an in-depth description of this source and its references the reader may refer to http://diss.rm.ingv.it/dissHTML/ITCS076INF.html.

Owing to their location and inferred hypocentral depth, infrequent events beneath the Po Plain may be associated with the reactivation of inherited faults. Among these, the 15 May 1951, M_w 5.4 *Lodigiano* earthquake (ID 8) occurred in a poorly investigated region between the most advanced fronts of the Northern Apennines and those of the Southern Alps, north of the Emilia Arc (Fig. 9a). Its location is made less certain by its unusual intensity pattern, which appears rather spread out, thus suggesting a deeper source (MELONI et al. 1988). Nevertheless, the 1951 event falls outside the two compressional systems and far from any of the transverse structures discussed in the next section (Fig. 3). For this reason we suggest that it may have been generated by one of the inherited faults of the foreland, even if its true seismogenic source has not been identified yet. The location of this earthquake was already reported as unusual by CALOI et al. (1956), who remarked that it falls in a totally aseismic region. Owing to the presence of several productive gas fields in the epicentral area, they concluded that the event could be seen as an induced earthquake, which would explain its unusual location. Based on its presumably large depth, however, and to the ongoing compression between the Emilia Arc and the Southern Alps, we rather interpret the 1951 earthquake as a natural event generated by one of the inherited faults that exist in the metamorphic basement (ROGLEDI 2010; Fig. 9a).

Two other candidate earthquakes for having been generated by inherited fault systems are the 22 October 1796, M_w 5.6 *Emilia orientale* and the 13

January 1909, M_w 5.5 *Bassa Padana* "deep" earthquakes (ID 15 and 16, respectively). These events are located close to each other within the Ferrara-Romagna Arc. As discussed in the following section, they occurred beneath the Po Plain at a hypothesized depth that causes them to fall below the basal detachment of the thrust wedge. As discussed below, their causative faults are still under investigation and may belong to Group #4.

3.5. Transverse Structures (Group #4)

The transverse structures are expressed as lineaments oriented mainly northeast-southwest in the Northern Apennines thrust belt, thus roughly perpendicular to the chain, and northwest-southeast across the Southern Alps. They were hypothesized as early as the 1950s based on geomorphological observations and on evident lithological and facies discontinuities in the regional geological texture (e.g. SIGNORINI 1935; BOCCALETTI et al. 1977; FAZZINI and GELMINI 1982 for the Apennines; CASTELLARIN and VAI 1986; CASTELLARIN et al. 1998, for the Alps). Although their nature and kinematics are still largely unknown, they seem to subdivide the Northern Apennines and Southern Alps thrust fronts into 30–50-km long sections (e.g. CASTELLARIN et al. 1985). In the first public release of the DISS database (v. 2.0, VALENSISE and PANTOSTI 2001b) they were compiled, georeferenced, and plotted along with known seismogenic sources and major earthquakes, suggesting (1) that they play a passive role in controlling the current distribution of seismic release, but also (2) that they may themselves host significant earthquakes. Unfortunately, there is little evidence for the exact geometry and the activity of these lineaments, both at the surface and in the subsurface, and even their role in the geodynamic framework of the Northern Apennines and Southern Alps is still the matter of debate (e.g. SORGI et al. 1998; ELTER et al. 2012).

Most of the CSSs discussed in this paper terminate where the most significant of such transverse structures bound palaeogeographic domains, effectively acting as boundaries to earthquake propagation. The CSSs of the Southern Alps are bounded to the east by the Schio-Vicenza Line (Fig. 7), a NW–SE, 120-km long, subvertical strike-

slip fault (e.g. PELLEGRINI 1988; CASTALDINI and PANIZZA 1991; VIGANÒ et al. 2008). The Schio-Vicenza Line does not appear to have generated earthquakes larger than magnitude 5.5, but significant earthquakes are expected along the neighboring CSSs of the Eastern Southern Alps, shown in blue in Fig. 1 (see GALADINI et al. 2005; BURRATO et al. 2008 for an overview).

We interpreted the transverse structures as the brittle response of the upper lithosphere to the lateral changes in the geometry of the subducting slab as reflected in the dip of the foreland monocline (MARIOTTI and DOGLIONI 2000). This interpretation is based on the observation that (1) the thrust belts show undulations controlled by inherited Mesozoic horsts and grabens (see Sect. 2.2), with recesses occurring around horsts and salients developing in grabens (Table 1), and that (2) the terminations of these undulations match the transverse structures and are reflected in the shape of the CSSs, resulting in first-order segmentation of the compressional fault systems.

The deeper earthquakes play a central, albeit poorly understood, role in the assessment of the seismogenic potential of these transverse structures. Over the past few decades, a significant amount of moment has been released in relatively deep earthquakes generated by unknown sources. Figure 9a shows the lineaments (i.e. transverse structures) reported in the literature, and the instrumental and historical earthquakes deeper than 15 km (see Tables 2, 3). The focal mechanisms of these deep earthquakes are both reverse and strike-slip. In detail:

- the source of the 13 September 1989, M_L 4.7 earthquake (ID 40), having a depth of 40 km and a strike-slip focal mechanism, falls below the Schio-Vicenza Line and is thus assigned to it;

- the majority of deep instrumental and historical earthquakes (ID 41, 45, 11, 62, 29, and 63) fall in a relatively small area of the Apennines and Pedeapennines. The 13 March 1832, M_w 5.5 (ID 11) and the 17 September 1873, M_w 5.4 (ID 29) earthquakes exhibit the characteristic signature of deep earthquakes (Fig. 5). The 2 May 1987, M_L 4.5 (ID 45) and the 9 November 1983, M_L 4.8 (ID 62; Fig. 5c) events are both reported at a depth around 25 km and exhibit strike-slip and transpressional

focal mechanisms, respectively. The 25 January 2012, M_L 5.0 (ID 41) and the 27 January 2012, M_L 5.2 (ID 63) earthquakes exhibit a depth of 29 and 72 km with transtensional and reverse kinematics, respectively. The location of all these events corresponds to the location of significant lateral changes in the foreland monocline. The monocline plunges beneath the Apennines thrust belts and accretionary wedges with a dip that varies laterally between 14°, 8°, and 23°. This area is a crucial crossroad as it controls the forward propagation of the shallow thrusts, ultimately causing a significant northward shift of the Po River near Guastalla. Furthermore, this area of deformation falls at the northern end of a major transverse structure, namely the Enza Line (EL in Fig. 9), a strike-slip fault inferred from stratigraphic and structural data displaying dip-slip kinematics in Middle-Late Pliocene–Pleistocene (CASTELLARIN et al. 1985);

- We propose that the source of 3 October 2012, M_L 4.5 earthquake (ID 61), having a depth around 30 km and a strike-slip focal mechanism, marks a less visible transverse structure located in between the Upper Val Trebbia and Stirone-Ongina Lines (CASTELLARIN et al. 1985).

- The 25 June 1869, M_w 5.4 earthquake (ID 30) is a deep event that occurred near the junction of the Apennines deep ramps and the Viareggio-Val di Lima-Bologna transverse structure (VVL in Fig. 9). As such, it might belong equally well to Groups #2 or #4.

- The 14 September 2003, M_w 5.3 earthquake (ID 66) already described in Group #2 makes an interesting case. With a depth of about 20 km and a reverse focal mechanism its mainshock is likely to have been generated by one of the deep ramps, but the subsequent aftershock sequence delineated likely a subvertical fault probably corresponding with the (PSL in Fig. 9). This transverse structure is inferred from remote sensing, surface geology and structural data (BOCCALETTI et al. 1985; CASTELLARIN et al. 1985).

- The 22 October 1796, M_w 5.6 (ID 15) and 13 January 1909, M_w 5.5 (ID 16; Fig. 5b) earthquakes exhibit the characteristic signature of the deep earthquakes, and occurred at approximately the

same location of the southeastern Po Plain (see Sect. 2.3). These two events are therefore below the thrusts of the Ferrara-Romagna arc. In principle they may have been caused by as yet unknown inherited structures of the Group #3. Their locations, however, fall near a strong lateral variation of the dip of the foreland monocline, possibly marking the extension of the (PSL in Fig. 9) beneath the Po Plain deposits. To the south, west of the Mugello basin, the same lineament appears to interrupt the continuity of the Etrurian Fault System, a major extensional system straddling the crest of the Northern Apennines (Fig. 9a). When placed in a palaeogeographic perspective, the Prato-Sillaro Line appears much as the mirror image to of the Enza Line, although less obvious. We conclude that the 1796 and 1909 earthquakes may have been generated by the northeastern extension of the PSL, and should thus belong to Group #4.

- The 5 April 2009, M_L 4.6 earthquake (ID 68), occurred with a transpressional focal mechanism below the shallow Pedeapenninic Thrust Front, but at a depth of about 25 km. Based on its location, this event could have been generated by the Montone Line (M_L in Fig. 9a), therefore, belonging to Group #4.

- Finally, the 6 June 2012, M_L 4.5 earthquake (ID 60) occurred in the Adriatic Sea, just a few kilometers offshore, at the eastern end of the Ferrara-Romagna arc. Its 26 km depth and the dominantly strike-slip mechanism suggest it may be evidence for a sort of "rebound" of the eastern end of the arc only a few days after the strong shocks of 20 and 29 May 2012.

In summary, a number of sizable historical and instrumental earthquakes of the Northern Apennines seem to follow well defined trends that correspond with the main mapped transverse lineaments. This circumstance is further stressed by the seismic release pattern delineated by the 2009-2013 sequences (Fig. 9b). Indeed these mostly shallow sequences occurred far from the inherited faults described above (belonging to Group #3), but seem to delineate well the pattern of the transverse structures. These structures also play an important passive role in the

propagation of ground shaking caused by significant deeper earthquakes (e.g. see the 25 January 2012 earthquake: http://mappe.haisentitoilterremoto.it/2221246460/index.html, or the 27 January 2012 earthquake: http://mappe.haisentitoilterremoto.it/2221279330/mcs.jpg, both of which propagated efficiently southeastward only up to the Prato-Sillaro Line. Similarly, the aftershocks of the 2012 Emilia sequence terminated against the Secchia Line (SL in Fig. 9b), which played on obvious passive role.

4. Discussion and Conclusions

4.1. Identifying Seismogenic Sources in the Po Plain

The correct identification of seismogenic sources in the Po Plain is a crucial and extremely complex task that requires unconventional data and innovative approaches and methodologies. As most of the seismogenic faults are blind and cannot be detected by standard field mapping, the vast majority of the relevant input must be derived from subsurface data, drainage and sedimentary patterns, and seismicity.

Despite these unfavorable circumstances, the investigation of the local seismotectonic setting is somewhat easier than in other parts of the Italian peninsula. The Po Plain has undergone steady compression and shortening from the Upper Pliocene to present, generating a relatively monotonous series of thrust faulting systems that are progressively younger as one moves from west to east and from the inner to the outer portions of the two confronting orogens, the Alps and the Apennines. In marked contrast, the inner portions of the Apennines have witnessed a recent and drastic change in the dominant stress field and the onset of significant regional uplift. The combination of these two circumstances makes field mapping especially difficult and potentially misleading, as the limited cumulated effects of the present-day extensional tectonics are masked by the occurrence of faults generated under previous regimes and rejuvenated by the ongoing uplift (VALENSISE and PANTOSTI 2001a).

Current slip rates are relatively low on faults of the Po Plain, implying more infrequent earthquakes and thus a lower seismic hazard. The analysis of

cumulative strain over different time intervals, however, shows that at least in the southern portion of the Po Plain deformation used to be faster but has steadily slowed down in recent geological times (MAESANO et al. submitted).

The correct estimation of slip rates requires proper considerations of the effects of differential compaction (SCROCCA et al. 2007; MAESANO et al. 2013). The cumulative shortening of the two confronting orogens has created significant relief in the subsurface, where carbonate ridges nearly reaching the topographic surface are juxtaposed to deep troughs filled by as many as 8,000 m of Pliocene–Pleistocene sediments. These deposits are naturally bound to reduce their volume; the expected volume reduction, however, is proportional to the thickness of the unconsolidated sedimentary layers, which varies greatly from place to place. As a result, the compaction will be differential, i.e. it will be more pronounced in the deep troughs than on top of the anticlines (e.g. MAESANO et al. 2013). The ensuing differential volume loss may simulate the outcomes of tectonic activity even over long time periods after activity has ceased. In other words, tectonic activity associated with crustal shortening promotes the creation of uplifted areas, where sedimentation is limited or absent, and lowlands that act as sediment traps, but differential compaction itself induces relative thickness changes of the sedimentary layers, simulating the geometry of growth strata and causing a trade-off between the outcomes of genuine tectonic activity and standard sedimentary effects. All in all, differential compaction may lead to a gross overestimation of the tectonic rates, in the milder cases, or to mistaking a dead for an active fault, in the most severe instances. Assessing the role of differential compaction is thus a crucial step in the assessment of the earthquake potential of faults buried by a thick blanket of unconsolidated sediments.

Accurate seismic hazard assessment in the Po Plain is important due to the large number of industries and critical infrastructures and to the concentration of over 10 million people, about 20 % of the country's population. And even though Po Plain earthquakes are generally quite moderate compared to those located in peninsular Italy, anticipating where, how frequently, and how strongly they

may strike takes on a special significance and requires special care. As in many other problematic areas worldwide, no individual dataset alone may allow us to identify all seismogenic sources unambiguously, and resorting to a blend of diverse observations becomes mandatory. Because of the limited geological strain rates and to the sparseness of seismicity, any conclusion must be carefully constrained by regional-scale geodynamic models and evidence for ongoing deformation (GPS). The seismotectonic model presented in this paper is indeed the result of such blending, and is also a scheme that will be easy to test as more data become available.

4.2. Observed and Expected Seismicity

The Po Plain is the locus of moderate-size (normally $<M$ 6.0) and infrequent earthquakes, at least compared to peninsular Italy, and probably for this reason its seismic hazard has long been under-recognized. Earthquake codes have acknowledged the local earthquake potential only since 2004, and it has taken six more years for the new provisions to be fully enforced. This has resulted in inadequate earthquake protection which largely explains the human and structural loss caused by the May 2012 Emilia earthquakes (M_L 5.9 and 5.8), when 27 people died in collapsed factory buildings (STUCCHI et al. 2012).

We subdivided the damaging Po Plain earthquakes into shallow and deep, depending on whether they occur above or below 15 km depth (Fig. 3; Tables 2, 3). Shallow events are more damaging and sometimes catastrophic, whereas deep events usually do not cause extreme damage, but are strongly felt over large areas (Fig. 5). In summary, the Po Plain is expected to experience (see Fig. 5; Table 5):

- shallow, damaging earthquakes, such as the 17 November 1570 *Ferrara* and the 20–29 May 2012 *Emilia* events, which are responsible for 84–96 % of the total moment release (depending on the time frame considered);
- deep earthquakes, such as the 13 January 1909 *Bassa Padana* and the 9 November 1983 *Parmense* events, which are responsible for 16 or 4 % of the total moment release (depending on the time frame considered);

- sequences with multiple mainshocks, such as the 1504–1505 *Bolognese*, the 1570 *Ferrara*, the 1781 *Faenza*, the 1916 *Rimini* and the 2012 *Emilia* events; in fact, the majority of crustal earthquakes, both historical and instrumental, exhibit this characteristic behavior (Table 4; Fig. 4); and

- infrequent destructive earthquakes, such as the 3 January 1117, M_w 6.7 *Veronese* event, that are difficult to anticipate as they are generated by the least known category of seismogenic sources of the Po Plain (inherited faults, Group #3).

Shallow earthquakes occur following a relatively regular pattern along major identified buried thrusts, with a concentration on the external front of the Ferrara-Romagna Arc and on the Pedeapenninic Thrust Front (Fig. 3). In contrast, deep earthquakes are concentrated in relatively small areas, possibly coincident with the transverse structures (Figs. 3b, 9a). All in all, the Northern Apennines structures (Emilia Arc, Ferrara-Romagna Arc, and Pedeapenninic Thrust Front) exhibit a larger seismogenic release compared to the Southern Alps (Southern Alps Outer Arc plus Southern Alps Mountain Front), both in historical times and instrumental. More specifically (Figs. 3, 7):

- the earthquake potential of the main fronts of the Northern Apennines increases from west (Monferrato Arc, Emilia Arc) to east (Ferrara-Romagna Arc);

- the Pedeapenninic Thrust Front, that lies close and beneath the main urban areas of Emilia-Romagna, has released significantly in historical times but has been almost totally quiescent in the past few decades;

- the deep ramps host a small fraction of the total earthquake potential;

- multiple sequences occur preferentially on shallow thrusts, and specifically on those belonging to the Emilia and Ferrara-Romagna Arcs.

Finally, Po Plain seismogenic sources exhibit a large variability in terms of maximum magnitude, ranging between 5.5 and 6.7 (Fig. 10). Sources expected to generate the strongest earthquakes include that responsible for 1117 earthquake (ITIS140, ITCS076) and the deep ramps (ITCS027).

Sources of the Ferrara-Romagna Arc, of the Pedeapenninic Thrust Front, and the thrusts and backthrust of the outer arc of the Southern Alps are known to have generated earthquakes up to M_w 6.1. Finally, sources of the Southern Alps Mountain Front, of the Giudicarie System and of the Emilia Arc have not exceeded M_w 5.7 historically.

4.3. Summary and Open Issues

We described a large set of seismogenic sources encompassing the Po Plain. The vast majority of the sources exhibit a minimum and maximum faulting depth of 1–5 and 7–10 km, respectively, reflecting the characteristic geometries and depths of the Alps and Apennines accretionary wedges. Exceptions are the deep ramps (ITCS027), where earthquakes are assumed to take place at a depth of 12–22 km, and the transverse structures, that appear to generate seismicity over a wide depth range.

Because of obvious geological reasons, the accuracy of our seismotectonic model decreases rapidly as faulting depth increases. If it may be reassuring to learn that most of the moment is released by shallower and thus easier to identify sources, we must keep in mind that they also cause the strongest ground shaking. In contrast, future earthquakes are likely to be spawned by deep and largely unexplored seismogenic sources, but we also know that the associated ground shaking will be milder. The investigation of these deep sources is a priority for future research efforts. These will necessarily include an improved consideration of the role exerted on thrust development by the palaeogeography of the Alpine margin, a clarification of the role of the transverse structures cross-cutting the Northern Apennines, and a reassessment of the seismotectonic evidence supplied by seismicity.

Perhaps the main obstacle towards a better understanding and more accurate assessment of the earthquake potential of the Po Plain, however, is the question of the limited *earthquake efficiency* or *seismic coupling* of the local seismogenic sources. BARBA *et al*. (2010) have shown that the seismic moment released by the compressional earthquake sources encircling the Po Plain is <25 % of the
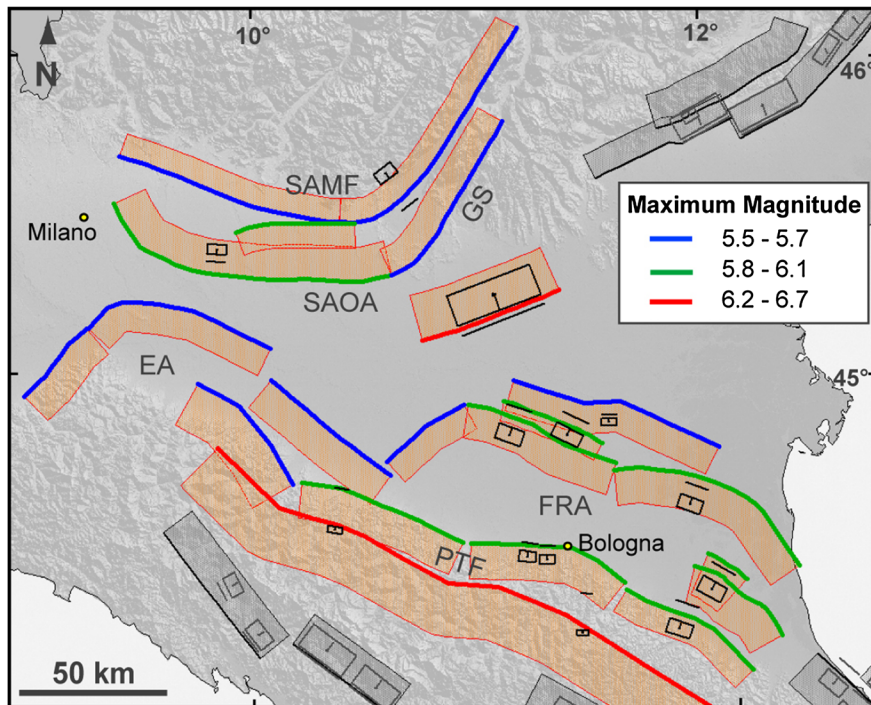
Figure 10
Maximum expected magnitude estimated for all the composite seismogenic sources from the database DISS discussed in this work

moment estimated from geological observations and geodynamic modeling. This brings up some questions: Is *low efficiency* a permanent characteristic of the Po Plain, or perhaps of all similar thrust faulting systems worldwide? Does *low efficiency* imply that the same faults behave seismically and aseismically at the same time, or does it mean that many faults slip aseismically without ever generating earthquakes while a few only slip coseismically? There may well be a deficit in historical moment release, but how accurately do we know the historical earthquake record? Can we rule out the occurrence of rare large earthquakes that would rebalance the earthquake budget?

Instrumental monitoring of the Po Plain has provided reliable earthquake locations and focal mechanisms over the past 30 years. In contrast, the historical seismicity record is likely to be incomplete for at least two reasons. The first and principal reason has to do with the environmental characteristics of large portions of the Po Plain. Unbeknown to many, this industrialized and heavily urbanized part of Italy was once covered by large forests and marshlands,

and thus sparsely populated. Before the construction of large modern levees, some of which were completed as late as the end of the seventeenth century, many Po Plain rivers meandered across the plain driven by the distribution of topographic highs and lows. The second reason is that much of the seismic moment may have been released in deep earthquakes, but since these cause only mild shaking many of such events may have gone undetected through history and until relatively recent times.

As a result, the pre-instrumental earthquake record may be locally incomplete, particularly in the marshy areas of the southern Po Plain between the Ferrara-Romagna Arc and the Pedeapenninic Thrust Front. Much of the available information refers to earthquakes recorded at the cultural and economic centers, such as Ferrara and Mirandola, normally located in dry spots on top of buried anticlines, making it difficult to separate small, local from larger and more distant events. This implies that the effective completeness of the Po Plain earthquake record may be as short as a few centuries, even where tectonic deformation is fastest such as in the central-

southern Po Plain, thus inducing and explaining the known fast landscape changes.

In summary, how accurate is the seismotectonic model described in this paper? The 20 and 29 May 2012 Emilia-Romagna earthquakes and the associated sequence indeed validated it, and the overall match between the observed seismicity and the current geological understanding is encouraging. Yet the biggest challenges may still lie ahead of us.

Acknowledgments

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